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N = notis. F = foredrag. FR = foredrag referert.
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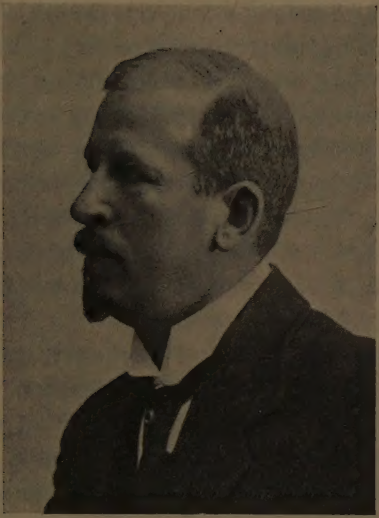

CARL FRED. KOLDERUP OG UTFORSKNINGEN AV VESTLANDETS GEOLOGI

AV

ANDERS KVALE

Vestlandet var i forrige århundre dårlig forsynt med karter, det var slitsomt å reise og bergartene var vrine, særlig leit var det at fossilførende bergarter syntes å mangle. Det var ikke å undres på at utforskningen av denne landsdel trådte i bakgrunnen i forhold til Østlandet og Trøndelag. Helt ukjent var dog Vestlandet ikke, det ble besøkt alt av Bedemar, v. Buch og Naumann — den siste oppdaget det vi nå kaller Bergensbuene — og de første norske geologer gjestet enten selv Vestlandet eller sendte sine assistenter dit på kortere turer. Men fram til 1860 var det praktisk talt ikke utkommet en eneste avhandling om Vestlandets geologi. På den tid begynte en og annen vestlending å interessere seg for geologi, og det falt nærmest av seg selv at de tok fatt på geologiske undersøkelser i hjemlige trakter. Bergenserne Hiortdahl og Irgens utgav i 1862 „Geologiske undersøgelser i Bergens omegn", og skrev senere om de geologiske forhold mellom Sognefjorden og Stadt. Hardingen S. A. Sexe skrev i 60- og 70-årene om kvartærgeologiske emner, særlig fra Vestlandet. Bergenserne Helland og Reusch utgav fra 70-årene og framover en rekke viktige arbeider, Helland særlig om praktisk-økonomiske emner, Reusch petrografiske og tektoniske arbeider. For vårt kjennskap til Vestlandets geologi var Reuschs arbeider de viktigste, særlig må nevnes hans hovedverker: „Silurfossiler og pressede konglomerater i Bergensskifrene“ (1882) og „Bømmeløen og Karmøen med omgivelser“ (1888), som begge var meget betydelige arbeider. Når vi tar med bergenseren T. Ch. Thomassens avhandling om Folgefonnshalvøya fra 1877, lite i omfang, men bygget på samvittighetsfult feltarbeide, har vi fått en oversikt over det som var gjort i Vestlandets geologiske utforskning før Kolderup.

Carl Fred. Kolderup ble født i Bergen 6. desember 1869. Han tok artium i 1888, gikk et år på krigsskolen, og ble cand. real. allerede i 1892 med geologi som hovedfag. Han var Brøggers første elev ved Universitetet i Oslo, og, som rimelig kan være, ble det petrografien, spesielt eruptivbergartene som fanget hans interesse. Kjennskapet til eruptivene undergikk i 90-årene en rivende utvikling etter at bruken av mikroskopet hadde slått igjennom i tiåret før, og Brøgger var ved siden av Rosenbusch den ledende petrograf i samtiden.

Som sin første spesialoppgave valgte Kolderup feltet av labradorsteiner og beslektede bergarter i Egersund—Sogndalstrakten. Det var ukjent land i dobbelt betydning, feltet var lite kjent, og det vi vil kalle anortosittstammen var lite utforsket. Feltet var på 1500 km^2, det topografiske underlag var amtskartet, veier var det dårlig med og reiseforholdene meget tungvinte. Hovedvekten måtte legges på å beskrive de enkelte bergarter og deres innbyrdes forhold. Avhandlingen utkom 1896 og fikk tittelen „Die Labradorfelse des westlichen Norwegens I“, en tittel som gav et vink om de oppgaver han alt da hadde satt seg. Dette arbeide var et betydelig bidrag til anortosittstammens petrografi og systematikk.

I 1896 ble Kolderup knyttet til Bergens Museum som styrer av de dengang meget beskjedne mineralogiske og geologiske samlinger. I de første år var det en bistilling, han arbeidet i skolen om formiddagen og på museet om ettermiddagen, men i 1904 ble han dosent og i 1914 professor. I og med ansettelsen ved museet var hans livsoppgave klar: å utforske Vestlandets geologi, og med denne oppgave arbeidet han utrettelig i over 40 år.

Det var naturlig at han begynte med labradorsteinene i Bergensfeltet. Han arbeidet med dem gjennom flere år ved siden av andre oppgaver. Labradorsteinene var nokså forskjellige fra dem i Egersundsfeltet, parallellstrukturen var meget framtredende, og spesielt de monzonittiske bergarter kunde ofte være vanskelige å skille fra gneiser. Hvis en sammenligner hans kart i „Die Labradorfelse des westlichen Norwegens II“ (1903) med kartet fra 1940, vil en finne betydelige avvikelser; men hans detaljerte og samvittighetsfulle bergartsbeskrivelser og hans studier av bergartenes innbyrdes forhold har beholdt sin verdi. Blant de viktigste resultater av dette arbeide kan nevnes hans oppdagelse av den monzonittiske bergart hvis mikroperthitt har spindler av oligoklas, og som han kalte mangeritt etter Manger i Nordhordland. Også denne bok var et betydelig arbeide, som sammen med „Die Labradorfelse I“ gjorde hans navn kjent over hele den geologiske verden. Et mål for den oppmerksomhet de vakte, er at hans vakre typesamlinger fra de to felter ble et fortrinlig byttemateriale, til stor nytte for museets dengang beskjedne geologiske samlinger.

Denne avhandling satte et foreløbig punktum for hans arbeide med labradorsteinene. Selv om de store felter i strøket Voss—Indre Sogn kunde friste, ventet det andre oppgaver som var viktigere. Det bør også nevnes at han i 1898 under tittelen „Lofotens og Vesterålens gabbrobergarter“ beskrev en samling prøver fra disse strøk.

Kolderups første offentliggjorte arbeide fra Bergenstraktene var: „Et orienterende nivaa i Bergensskifrene“ (1897). Et funn av fossilførende kalksten i Samnanger skapte tryggere grunnlag for parallellisering av Samnangerfeltets og Osfeltets bergarter og dermed for utredning av forholdene i den ytre Bergensbue. I 1901 utkom „Fjeldbygningen og bergarterne ved Bergen“ av Reusch og Kolderup, hvor Reusch skrev den geologiske del og Kolderup den petrografiske. Dette var hans første befatning med Vestlandets metamorfe bergarter.

Etter avslutningen av labradorsteinarbeidet tok han, inspirert av Brøggers store arbeide over de senglaciale og postglaciale nivåforandringer i Oslofeltet, fatt på en undersøkelse av de kvartærgeologiske forhold i Bergensfeltet. Her var nytt land å rydde, og i de følgende år fikk han utført et betydelig arbeide med oppmåling av terrasser og strandlinjer, undersøkelser av grustak og innsamling og bestemmelse av skjellprøver. Resultatene ble samlet i „Bergensfeltet og tilstøtende trakter i senglacial og postglacial tid“, som utkom i 1907 og som innbragte ham den filosofiske doktorgrad. Kvartærgeologien har på mange måter gjennomgått en rask utvikling siden dengang, og det gjelder for denne bok som for dens forbillede, at enkelte av de oppfatninger som ble hevdet vil måtte modifiseres; men den vil bli stående i framtiden som en viktig materialsamling for utforskningen av Vestlandets kvartærgeologi.

Etter denne visitt til kvartærgeologien vendte Kolderup seg igjen til petrografien, og tok nå fatt på den oppgave som ingen geolog på Vestlandet kommer utenom: de metamorfe bergarter. Hans utdannelse hadde ikke gitt ham et så solid grunnlag for studiet av disse bergarter som av eruptivene, heller ikke for tektoniske undersøkelser i fjellkjedestrøk. Oppgaven var derfor vanskeligere enn dem han før hadde stilt seg, men han tok fatt med sin vanlige energi. Han begynte med en detaljert undersøkelse av Samnangerfeltet, en del av den ytre Bergensbue, i nordlig fortsettelse av Osfeltet, som var kjent gjennom Reuschs arbeide fra 1882. Bergartene var delvis de samme, men de var sterkere omvandlet. For å skaffe et skikkelig topografisk underlag for kartlegningen, som foregikk i 1 : 25 000, måtte han utføre betydelige korreksjoner på rektangelkartet. Alt i alt tok det fem år før dette felt var ferdig bearbeidet. Resultatet ble en større avhandling: „Fjeldbygningen i strøket mellem Sørfjorden og Samnangerfjorden i Bergensfeltet“, som utkom 1914—15. Avhandlingen er utpreget deskriptiv, med utførlige og detaljerte beskrivelser av de opptredende bergarter. Hvor det kunde være tvil om deres opprinnelse, var han forsiktig i sine slutninger. En rekke kjemiske analyser øker bokens verdi. I bearbeidelsen av disse og i systematikken for øvrig fulgte han Grubenmann. Blant de mange viktige undersøkelser i dette arbeide må nevnes hans detaljerte studier over olivinsteinens forandring til serpentin.

Mens arbeidet med Samnangerfeltet pågikk, hadde han også andre jern i ilden. I 1911 utgav han sammen med sin assistent, senere lensmann P. O. Ottesen, en avhandling om Utsiras fjellbygning og bergarter. Samme år kom et arbeide om de såkalte hvite granitter i Sogneskollen ved munningen av Sognefjorden og i Bremangerlandet i ytre Nordfjord. Han påviste at de sto nær de amerikanske granodioritter i mineralogisk og kjemisk sammensetning. Et tredje arbeide fra samme år var en oversikt over Bergensfeltets geologi skrevet for en større ekskursjon av engelske geologer. I 1914 utkom en beskrivelse av kartbladet Egersund, hvor han med det nye rektangelkart som grunnlag hadde supplert sine undersøkelser fra 90-årene.

Etter avslutningen av Samnanger-arbeidet satte han seg en ny stor oppgave: å utforske konglomerat- og sandsteinsfeltene i Sogn og Fjordane fylke. Disse var delvis kjent fra Hellands og Reusch's arbeider, men hadde fått øket interesse ved de funn av devoniske plantefossiler som ble gjort fra 1902 og utover, dels av Kolderup, dels av hans elever og assistenter. Beskrivelsen av det minst kjente felt, Bulandets og Værlandets felt, kom alt i 1916. Det gikk fem år før den neste kom, og det hadde sine grunner. I 1914 var Kolderup blitt museets direktør, samtidig deltok han aktivt i det politiske liv og var fungerende ordfører under og etter Bergensbrannen. Han måtte således samtidig være med og planlegge museets utvidelse til universitetet, et arbeide som gjorde store framskritt i krigsårene, og byens gjenreisning etter brannen. Det er i virkeligheten forbausende at han i det hele tatt fikk tid til vitenskapelig arbeide i disse årene. Fra 1921 kom imidlertid beskrivelsene av devonfeltene slag i slag, det sjette og syvende i 1926. Disse arbeider gir en samlet oversikt over det som var mulig å finne ut om Vest-Norges devon med det nåværende kartunderlag. Ved siden av detaljerte beskrivelser ble framlagt en rekke forhold av generell interesse. Foruten funnene av plantefossiler, som tillot en sikker aldersbestemmelse, kan nevnes at sandsteinslagene til dels var foldet, at devonen til dels var skjøvet over sitt underlag, at det etter lagenes avsetning var foregått forkastninger, og overskyvninger av kaledonske bergarter over devonen, og at sure intrusiver hadde trengt opp i devonlagene. Disse trekk betyr en viktig utdypelse av vårt kjennskap til den kaledonske orogeneses siste faser i Norge.

Allerede før devonmaterialet var ferdig bearbeidet tok han fatt på den oppgave som han anså for sin viktigste, og som han hadde samlet materiale til hele tiden siden han kom til Bergen, nemlig å gi en oversikt over Bergensfeltets geologi. De innviklete petrografiske og tektoniske forhold gjorde oppgaven vanskeligere enn noen av dem han tidligere hadde løst; men med en menneskealders forskerarbeide bak seg sto han godt rustet til å ta fatt. De prektige Bergen Umland kartene som ble opptatt i begynnelsen av 20-årene, gjorde det mulig å foreta meget detaljerte undersøkelser, og ved hjelp av en rekke assistenter ble alle 8 blad geologisk kartlagt. Bearbeidelsen av det veldige materiale ble en del sinket av de mange andre gjøremål som la beslag på hans tid; men han hadde fått gjort unna atskillig da hans helse i midten av 30-årene tok til å svikte. Han hadde hatt kjempekrefter, men et langt liv i ustanselig arbeide, uten ferier og nesten uten fritid ble til slutt for meget. Det ble klart at han ikke vilde makte å fullføre verket alene, og hans sønn, Niels-Henr. Kolderup, trådte inn som medarbeider. I 1940 var monografien ferdig og ble trykt i Bergens Museums Skrifter. Det ble et helstøpt verk med en meget konsentrert framstilling. I sin knappe form gir det et levende inntrykk av hvor komplisert Bergensfeltet er i geologisk henseende, hvilken rikdom av petrografiske og tektoniske problemer det byr på, hvilket veldig arbeide som her er nedlagt, og samtidig viser det at her er oppgaver for fortsatte undersøkelser i framtiden. Foruten den kjærkomne beskrivelse av Bergensfeltet gir avhandlingen en rekke nyoppdagelser; særlig må nevnes påvisningen av at de mystiske gneiser i „Ulrikkens gneisfelt" er sure og intermediære bergarter av Bergen-Jotunstammen i forskjellige metamorfe fasies.

„Geology of the Bergen Arc System" danner en vakker og verdig avslutning på et langt livs utrettelig forskerarbeide, og det vil bli stående som et av de store arbeider i norsk geologisk litteratur. Da det høsten 1940 innbragte forfatterne den Bergenske prisbelønning for vitenskapelig arbeide, var det bare en mening om at den var vel fortjent.

Carl Fred. Kolderups geologiske forfatterskap omfatter 34 avhandlinger og ca. 30 artikler, dertil kommer 29 avhandlinger om jordskjelv. Til sammen er dette ikke så lite for en enkelt mann, selv om det er fordelt på 45 år. Men dette er bare en del av hans livsverk. For å få den fulle forståelse av hva han utrettet, måtte en også behandle hans arbeide med å skape og styre mineralogisk geologisk institutt og samlinger ved Bergens Museum, hans arbeide som Bergens Museums direktør gjennom 25 år, som kommunemann og politiker, som medlem av en lang rekke offentlige styrer og komiteer. Hvis en er klar over for en utrolig tid alt dette ekstra arbeide tok, blir hans vitenskapelige produksjon en prestasjon av rang. Den var kun mulig ved at han arbeidet ustanselig, søndag som hverdag. Ferie var for ham et ukjent begrep. Selv når han en sjelden gang måtte holde sengen noen dager, måtte han ha sine bøker og skrivesaker for hånden. Uttrykket arbeidsjern har vært brukt om så mange. De som fikk høve til å lære Carl Fred. Kolderup å kjenne, vet at det vil være meget vanskelig å finne noen som uttrykket passer bedre på enn ham.

Det som særpreget Carl Fred. Kolderup som vitenskapsmann, var først og framst grundighet og nøkternhet. Han beskrev samvittighetsfult det han så, og trakk de slutninger som kunne trekkes på sikkert grunnlag; men var det rom for flere tolkninger, var han oppmerksom på dette og uttrykte seg med forsiktighet. Han forsøkte ikke å dekke over vanskelighetene eller å gi det utseende av at han hadde løst alle problemer. Han var således en typisk deskriptiv forsker, og det er dem som trenges mest når ukjent land skal undersøkes. Hans navn er ikke knyttet til dristige teorier og fantasifulle tolkninger; men han har samlet inn og tilrettelagt et veldig materiale om Vestlandets geologiske forhold, et materiale som vil beholde sin verdi uavhengig av skiftende hypoteser. Hans navn vil minnes så lenge det blir drevet geologisk forskning på Vestlandet.

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Trykt 14. februar 1945.

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STRUCTURAL PETROLOGY OF THE BYGDIN CONGLOMERATE

BY

Trygve Strand

With 3 figures in the text and 24 diagrams.

Abstract. The Bygdin conglomerate, a quartzite conglomerate in the Valdres Sparagmite, was deformed by an overriding thrust-sheet of igneous rocks moving towards SE. The boulders of the conglomerate were drawn out in a NW—SE direction. Accompanying this deformation there is folding on NW—SE axes, both stretching direction and axes of folding pitching commonly 10—20° NW. The fabric patterns of quartz and sericite in the deformed boulders are girdles with a direction NE—SW. The folding and the production of the girdles are interpreted as produced simultaneously with the mass transport towards SE, caused by a triaxial deformation by which shearing stresses operated on planes striking NW—SE as well as on planes striking NE—SW.

The Bygdin conglomerate, a brief name for it not intended to be introduced as a new stratigraphical term, is the quartzite conglomerate in the Valdres Sparagmite as exposed near the road at the eastern end of Lake Bygdin, about 1100 metres a. s. l., district of Valdres, Southern Norway, geographical position about 61° 19′ N, 8° 50′ E (see map Fig. 1). For further topographical orientation the map Slidre (E 31 aust) of the Ordnance Survey of Norway may be used, the same sheet is at present in preparation for publication as a geological map by the Geological Survey of Norway.

In the region at the eastern end of Lake Bygdin and environs the following rock units are recognized, of which only 3. and 4. occur in the area of map Fig. 1:

1. Archaean rocks forming the basement are exposed in a window of erosion about 10 km to the south.

2. The sedimentary cover of this basement is made out of phyllites with some intercalated beds of quartzite, being marine deposits of Cambrian and Ordovician age.

3. Resting upon these marine deposits is the Valdres Sparagmite, of which the Bygdin conglomerate is the upper member preserved in this tract. The Valdres Sparagmite consists for the greater part

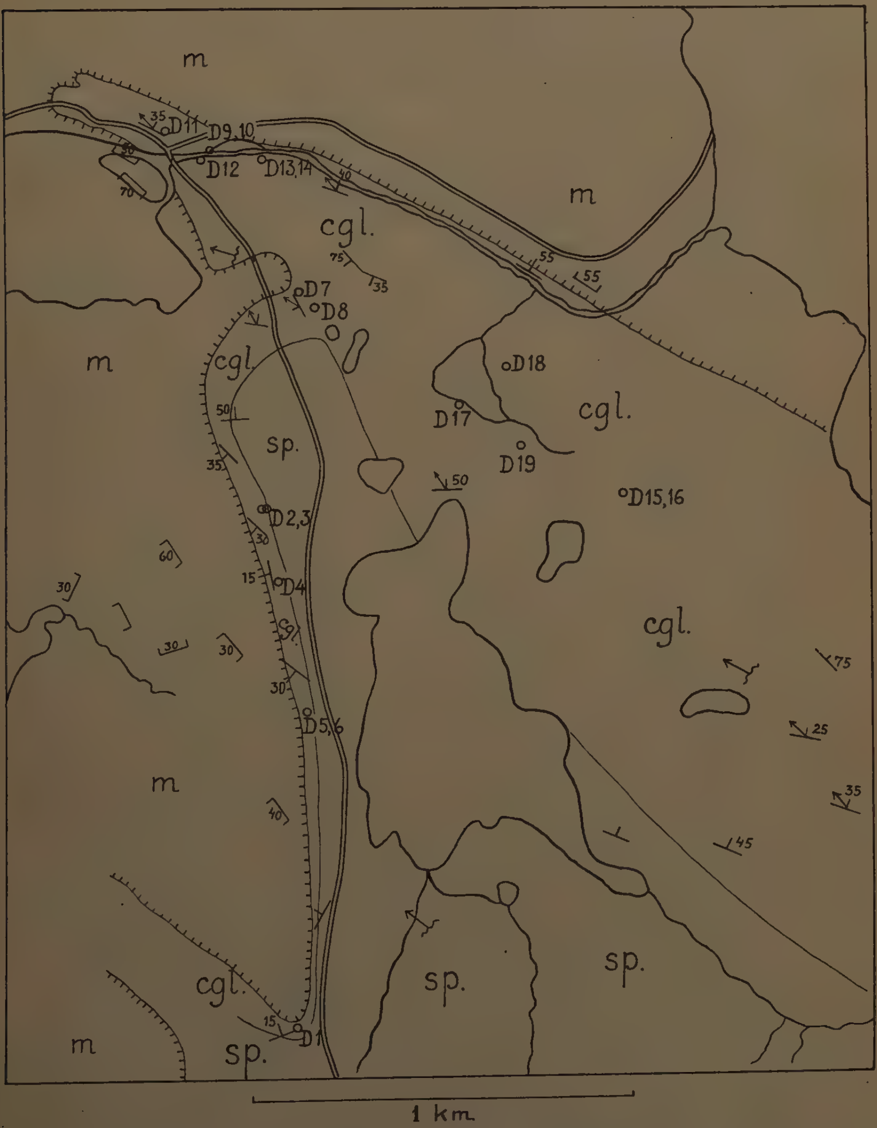

Fig. 1. Geological sketch-map of the quartzite conglomerate at the eastern end of Lake Bygdin (in the upper left corner). sp. = sparagmite, cgl. = quartzite conglomerate, m = mylonites. Arrows indicate pitch of the stretching direction or axes of small-scale folding (wavy lines). Rings indicate localities from which the specimens yielding the diagrams were collected.

of feldspathic arkoses, petrographically similar to the Eocambrian sparagmites. As has been definitely shown by Goldschmidt, however, it is of Caledonian age and is to be considered as a Caledonian flysch, or, perhaps better, molasse deposit (see Goldschmidt 1916 a, Strand 1939).

4. Overlying the Valdres Sparagmite with tectonical contacts is an overthrust massif of igneous rocks, out of which the Jotunheim mountains to the north of our area have been carved. For an account of these rocks, known as the Bergen—Jotun kindred, see Goldschmidt 1916 b. In the area of map Fig. 1 they are present as mylonites near the thrust-plane.

The Bygdin conglomerate was described in considerable detail by V. M. Goldschmidt in a well known paper (Goldschmidt 1916 a), in which was paid a special attention to the deformation of the conglomerate. A number of large specimens of the conglomerate collected by Goldschmidt are exhibited in the Geological Museum in Oslo and photographs are reproduced in Pl.s 2 and 3 in Goldschmidt's paper.

H. Cloos in his text-book "Einführung in die Geologie" has published a couple of photographs from the Bygdin area, one of which is of the deformed conglomerate.

The present writer collected a number of orientated specimens of the conglomerate in the area of map Fig. 1 and made rather detailed observations of the tectonics of the same area during visits in the summers of 1936, 1938, and 1939. The measurement in the laboratory of the quartz and sericite orientation was performed by a Leitz universal stage. The working out of the diagrams of the finely granulated quartz boulders was made possible by the use of special high power objectives for the universal stage from the firm Leitz (magnification 20 and 30 diametres). For the determination of the position of quartz axes the method of Tsuboi was used and found very convenient.[1] The diagrams were counted by means of 1 per cent circles. All diagrams here presented have been rotated into the horizontal plane. This procedure greatly facilitates the orientation for the reader and ought to be followed by all writers dealing with petrofabrics in connection with regional tectonics.

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[1] The method has been described in the Jap. Journ. of Geol. and Geogr., but volume and page of the publication can not be given due to the inaccessibility of the library of the Geological Survey during the present circumstances.

In the area of map Fig. 1 the structure is dominated by an anticline with axis trending NW, and pitching in the same direction. Erosion has brought up Valdres Sparagmite in the midst of the anticline with the overlying quartzite conglomerate at the sides, which is again overlain by overthrust mylonitized igneous rocks. As will appear from a glance at the map, the anticline is far from symmetrical, as the conglomerate occupies a wider space at the northeastern than at the southwestern side. At the southwestern side the conglomerate occupies a thin zone between sparagmite and overlying mylonites, commonly with a gentle dip in a western direction, while at the northeastern side it occupies a wide area, mostly with rather steep dips to the northeast. A calculation of the thickness from dip and extension across the strike would here give a great thickness, several hundred metres, for the conglomerate. From observations in regions farther east it is shown that the thickness of the conglomerate is less (about one hundred metres). The schistosity of the conglomerate, determined by the plane of the two greater dimensions of the deformed boulders, is here apparently not coincident with the stratification. This is directly seen in some outcrops showing the boundary between conglomerate and sparagmite with traces of stratification. Possibly the conglomerate is here also repeated by isoclinal folds.

The main features of the tectonics are seen from the map and from D 23, in which poles of bedding (and schistosity) planes in sparagmite and conglomerate have been plotted down together with the stretching directions (directions of elongation in the deformed boulders). As will appear from the diagram the poles of the bedding planes gather on a great circle, the pole of which coincides with the most common direction for the stretching, *i. e.* the stretching direction is coincident with the axis of folding. Another feature apparent in the diagram is the strong prevalence of northeasterly dip, the folds are commonly overturned to the southwest. In some localities one can also observe folding on a small scale with the same northwest axial pitch, which is quite symmetrical, with a sinus-curve shape of the folds. On cross-joints (normal to the stretching direction) in deformed conglomerates an other type of small scale folding can be observed in some places. In this case the boulders are bent and twisted about the stretching direction as an axis. The two minor dimensions of the boulders have no common orientation and the conglomerate has no schistosity (Fig. 2). This is clearly a shear folding (Scherfaltung,

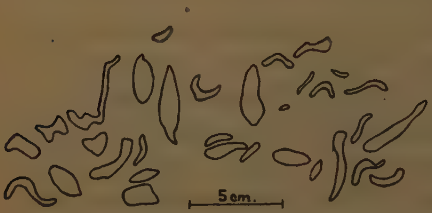

Fig. 2. Sections of deformed boulders on a cross-joint, showing shear-folding. Drawn from field sketches and a photograph.

Cloos) and must be caused by shearing stresses acting on planes parallel to the stretching direction.

In D 24 measurements of joints are represented by the poles of joints-planes. The joints fall on three groups. In the SE and also in the NW quadrant, near the horizontal plane, there are cross-joints or tension-joints normal to the stretching direction, often represented by two sets intersecting at a small angle. These are the commonest and most conspicuous of all joints, but they may be represented by relatively few measurements in the diagram. N—S striking joints with a steep dip to E, with their poles in the E quadrant, are interpreted as vertical shear-joints. Corresponding E—W striking joints, as demanded by symmetry, have not been observed. Finally, there are WNW—ESE striking joints, with the poles situated on a great circle, with a maximum concentration near the horizontal plane, representing steeply dipping planes. Also this group is interpreted as representing shear-joints.

Passing to the petrography of the conglomerate, the boulders are almost exclusively of fine-grained, flinty quartzites, with a grey colour of a greenish or sometimes reddish hue. Observations on undeformed conglomerates in other localities (Goldschmidt) show that the boulders were of sandstone with a well preserved clastic structure at the embedding in the conglomerate, so that their present structure is wholly a product of the deformation. Microscopically the structure of the boulders is characterized by very fine grain, the grains with irregular boundaries and marked undulous extinction. The matrix in the conglomerate is in most cases sparagmitic. According to Goldschmidt the matrix is remarkably less deformd than the boulders, showing clastic grains of feldspar in the microscope.

The appearance of the conglomerate is seen from Fig. 3. The form of the deformed boulders may be rod-like in extreme cases, but is generally as a triaxial ellipsoid. It may be remarked that the cross-section is more lenticular than elliptical. It is never possible to measure all three dimensions in one boulder, by combination of measurements of the major and medial axes on schistosity planes with measurements of the medial and minor axes on cross-joints, the following figures for the dimensions of boulders (in cm) were found by the author from the top of Svarthammeren east of Bygdin:

a	b	c	a : b : c	$d = \sqrt[3]{abc}$
39	12	7	5.6 : 1.7 : 1	14.9
37	9	6	6.2 : 1.5 : 1	12.6
28	9	6	4.7 : 1.5 : 1	11.5
18	8	5	3.6 : 1.6 : 1	9.0
15	4	1	15.0 : 4.0 : 1	3.9
13	4	1	13.0 : 4.0 : 1	3.7

Calculations of the original diameter, $\sqrt[3]{abc}$, of the boulders give an approximate measure of the changes which have taken place in the three dimensions, assuming a subspherical shape of the original boulders. Observations on undeformed conglomerate (in the locality Grønnsennknipa, Goldschmidt 1916a, p. 30) show that the boulders are really well rounded. The figures given above ought to indicate a diminuation of both the lesser axes, b and c, of the boulders, except for the smaller ones, in which b ought to be practically equal to the original diameter. The type of deformation characterized above is the common one, generally found in the area of map Fig. 1 as well as in other occurrences of the same conglomerate. For the Bygdin area Goldschmidt states the most frequent proportion of a : b : c as 6 : 2 : 1 (the mean of the above measurements is 5.8 : 1.8 : 1). In some cases the lengthening may be extreme, according to the same author up to 80 : 1.5 : 1.

A different type of deformation is represented by the so called "kvartskakelag" ("quartz-cake layers"), in which the boulders are extremely flattened, but where the difference between the two major dimensions is not extreme. Measurements of the two minor axes, b and c, from a conglomerate deformed in that manner from the area of map Fig. 1 are (in cm): 15—1, 11—0.8, 11.5—0.6, 9—0.4. Measurements of the corresponding major axes, a, could not be undertaken, but in this case there has certainly been a considerable widening of the dimensions b. The dimensions a are not so large as they would have been, if no such widening had taken place.

The orientation of quartz and sericite in the deformed quartzite boulders of the conglomerate will appear from the diagrams (D 1—20).

D 1—6 are from boulders deformed to "quartz-cakes" as described above.

Fig. 3. Surface of a cross-joint in the conglomerate, locality of D 12.

D 1 is quite unique in showing a well defined girdle with two maxima in the direction NW—SE. The specimen yielding the diagram was taken from the very contact between conglomerate and overlying mylonitic igneous rocks.

D 2, 3, 5 show distinct girdles with the direction NE—SW, as is the normal condition in the area. The girdles have two maxima, which in D 3 are bent down towards the horizontal plane and thus united into one. D 4 is quite atypical; if an error in the orientation of the section has not taken place.

D 6 shows a very distinct girdle orientation of the poles of sericite flakes, which is quite in accordance with the orientation of quartz axes. The same is the case with D 10, 14, 16.

D 7—19 are from quartzite boulders deformed in the common manner with a proportion between the two smaller dimensions about 2 : 1.

In D 7, 8 the girdles are asymmetric with a strong maximum in the NE part. D 9, 11 ,12 show less distinct girdles with asymmetry in the development of the maxima, and there is a tendency, apparent in D 11, of the strong maximum wandering towards the N quadrant. In D 13, 15, 17, 18 there are girdles, in which the two maxima apparent in other diagrams are united into one centering about the horizontal plane, while the zenith part of the girdles are sparsely set.

The maxima are also widened and drawn out in a horizontal direction. This has gone still further in D 19, where a horizontal girdle may said to be present. The stretching direction has here an uncommonly steep pitch.

D 20—22 are from sparagmites south of the area of map Fig. 1. These are reproduced to show that the orientation is much less sharp in these rocks, in which quartz is relatively subordinate to feldspar, mica and other minerals. They show the influence of "fabric neighbours" (Gefügegenossen) to lessen the effect of the orientation of quartz.

The salient feature in all diagrams, except of D 1, is the presence of a girdle with direction NE—SW, the axis of which, directed NW—SE, is parallel to the stretching direction of the boulders. In all diagrams there are maxima situated at a distance of 30° or less from the horizontal plane, while the zenith part of the girdles are always more or less sparsely set with quartz axez. The two maxima may have nearly symmetrical strength and position, giving the fabric a plane of symmetry, but the symmetry is never perfect and in many cases asymmetry is marked.

It is a noteworthy fact that the fabric patterns have a constant position in space, relative to the geographical coordinates, and that they, on the other hand, are independent of the s-planes. A fact worth attention is also the uniformity of diagrams taken from nearseated localities with the same type of deformation. This is seen from D 2—6 and from D 9—12.

No discussion of the mechanism active in the production of fabric patterns of the type presented will be taken up here. As long as the orientation mechanism of quartz is not more precisely known, such discussion will be of little value for the special problems here considered. So much is clear, however, that the girdles have been produced by partial movements in the rocks caused by shearing stresses, the partial movements as well as the stresses acting on planes situated in a zone, the axis of which is coincident with the axis of the girdle.

If the tectonics of the present area were to be interpreted from the facts presented above only, the result would most probably be the conception of a compression of the area in a NE—SW direction, in accordance with the trend of folding axes and the direction of the fabric girdles. Yet, in the areas of the Caledonian zone of deformation, of which the present area forms a small part, the direction of tectonical transport has been towards SE, thus the overthrust massif of igneous rocks above the deformed quartzite conglomerate has moved in this direction. We accordingly have folding axes in the direction of tectonical transport, and this is by no means particular of the present area. From many parts of the Caledonian mountain-chain of Scandinavia transverse folding (cross-folding) with an axial direction NW — SE to W—E has been noticed and often referred to in the literature. This folding commonly occurs together with "normal" folding with an axial direction NE—SW to N—S, a fine example is afforded by the Tysfjord area (Foslie 1941 with map), or it may be dominating over relatively large areas. Th. Vogt (1927, p. 167) also refers to a stretching structure with the direction W—E, coincident with the direction of thrust movement, from the Sulitelma area.

The transverse folding will necessarily have to be interpreted in one of two ways. Either it belongs to a separate act of deformation, older or younger than the normal folding, or both sets of folding are essentially simultaneous, and are expressions of one set of movement and deformation. The writer will emphasize two reasons against the first alternative. Firstly, from *a priori* considerations, it may be deemed wholly improbable that a mountain-making process should also produce active movements in a direction normal to the chief direction of mass transport. Experience shows, quite contrary, that foldings in the same area separated in time tend to follow the same trend-lines. Secondly, as the writer can testify from his own field experience, the two types of folding are intimately connected and can supersede each other within the narrowest limits. An example: On a bedding surface dipping NW and thus affected by the normal folding there may be quite small folds with axes pitching NW, representing the transverse folding. Then some metres away, the folds with NW-pitching axes have increased in dimension to a width, say, of one metre. The NW folding is then dominating and has obliterated the other folding.

It is accepted that rock deformation is generally biaxial or plane, that is the partial movements take place in the plane containing the direction of tectonical transport, the ac-plane of Sander. Normal to this plane, in the direction of tectonical strike, there is no movement, and any dimension in this direction remains unchanged. From a kinematical point of view this is simply ascribed to the fact that a rock mass has no freedom to move in the strike direction. Dynamically the matter is less simple, because this type of deformation requires a definite adjustment of the stresses. In this case shearing stresses will operate only on planes in a zone with the tectonical strike, b, as axis. The dynamical point of view in the treatment of rock deformation has been used by Walter Schmidt in his "Tektonik und Verformungslehre", in which some useful conceptions from technical mechanics are introduced into the discussion of rock deformation. Schmidt devotes a chapter in his book to the discussion of triaxial deformation (pp. 60—72), and arrives at the result that this type of deformation may be realized in some folding zones, in which movement in the general direction of strike has taken place. In this case the special adjustment of stresses leading to plane (biaxial) deformation has not taken place and shearing stresses will operate also on planes outside the zone of b. To take the area here dealt with as an example, there will be shearing stresses not only on planes striking NE—SW, but also on planes striking NW—SE.

The writer interpretes the folding with NW axes in the present area as due to a triaxial deformation of the rocks. From the above description of conglomerate deformed to "quartz-cake layers" it follows that in some cases elongation in the direction b has taken place. (The medial axis of the boulders, b, is in most cases coincident with the tectonical strike direction) Of course it is not possible for a rock mass as a whole to undergo any elongation in the main direction of strike, and an elongation that takes place in this direction in some parts of the mass must be compensated by folding and possibly also by deformation with shortening in the same direction in other parts of the mass. The transverse folding thus produced ought to be symmetrical, a condition which is not always fullfilled in our area, as the folds are often overturned to the SW, that is to the right of the direction of movement. The same is the case in the Tysfjord area, where an overturning of the transverse W—E striking folds, when present, always takes place to the south (Foslie 1941, p. 54). The Coriolis force might give an explanation of this circumstance, provided it should be of any concern at the very small velocities in rock deformation.

An example in miniature of folding with axes in the direction of the thrust movement has been described by Balk (1936, p. 738) in a thrust-faulted block of gneiss. The folds are subordinate features, but may attain several feet in length. There is some analogy between the miniature example described by Balk and the transverse folding in the Scandinavian Caledonian. The latter region is, indeed, built up of a series of very large thrust blocks, moving towards the foreland on a peneplained surface with a cover consisting mostly of argillaceous, incompetent rocks. It seems that these are the conditions, in favour of transverse folding. On the other hand, a thrust sheet is not supposed to be deformed or folded in a direction transverse to the direction of transport, as long as it is on its way upwards and is still inclosed in the depths of the earth (Schmidt 1932, p. 63).

The chief movement and deformation in the present area has taken place in the direction NW to SE, but in the particular area here considered, there is no folding with axes normal to this direction. Possibly the deformation producing elongation of boulders and pitch of the stretching direction and axes of folding commonly 10—20° NW was brought about by shearing on a single set of planes (einscharige Gleitung of Schmidt). Under the assumption of horizontal shearing planes, the pitch of the longer axis of the deformed boulders may be calculated from the axial ratio of the strain ellipsoid, as represented by the deformed boulders, by using the formulas of Becker (1893, p. 25) for this type of deformation. With the ratio 6:2:1 for the dimensions of the boulders, the direction of elongation of the boulders should make an angle of about 20° with the horizontal plane. With increasing deformation this angle will decrease. It is seen that this hypothesis is so far in accordance with observations. It may, however, also be argued that the regular girdle in D 1 indicates the simultaneous activity of both directions of maximum shearing stress in the ac-plane. In this case the strain should be essentially an irrotational one, and the axes of the deformed boulders should be coincident with the three principal stresses.

The prevailing fabric pattern in our area is correlated to the folding on NW—SE axes, parallel to the axes of the girdles. This folding, as well as the production of the girdles in the fabric patterns, was certainly brought about by shearing stresses acting on at least two planes normal to each other (mehrscharige Gleitung of Schmidt). The production of the girdles is so far easily interpreted. It is more remarkable that the partial movements in the plane of mass transport (the ac-plane) have left no records in most fabric patterns. (Such movement is recorded only in D 1.) It may be that the orientation produced by this movement was an unstable one in relation to the prevailing orientation on NE—SW girdles, or these girdles were formed in the last phase of the deformation and have thus been preserved in the record.

The chief feature of the diagrams here presented is the presence of girdles in a direction normal to the chief direction of movement, and, as interpreted by the writer, the partial movements producing the girdles and the main mass transport were integral parts of the same act of deformation. A similar case was recorded by H. Martin (1935) in diagrams from fault planes, in which the direction of movement was determined from field evidence. Sander refused to accept the correlation of the production of a girdle to a movement in a direction parallel to its axis, as shown by a citation from his polemics with Martin (1936): "Daß etwa, zuordenbar einer Verschiebung parallel einer Richtung n, Gürtel ⊥ n entstehen können, sozusagen als ein neuer Tektonit-Typus unter bisher nicht begegneten Bedingungen, — — —, das ist an Hand der von Martin vorgelegten Daten nicht wahrscheinlich zu machen." The present writer must agree with Martin that girdles are generally ambiguous as to the determination of the direction of movement.[1]

Thus in Sander's opinion a girdle is always a testimony of a separate act of deformation with a direction parallel to the girdle (normal to its axis). F. C. Phillips accepted this view in the interpretation of the fabric patterns of the Moine schists of Scotland worked out by him (1937). In the Moine schists, SE of the great Moine thrust, he showed the presence in the fabric pattern of quartz and mica of girdles with the direction NE—SW. The axes of the girdles are coincident with a lineation or stretching structure with the direction about NW—SE and the pitch SE. As the Moine thrust is directed towards NW, there is consequently a full analogy with the area here described, the opposite direction of pitch being in accordance with the opposite direction of movement. In the interpretation by Phillips the lineation structure and fabric girdles had their origin in a deformation act with a compression in the direction NE—SW prior to the Caledonian thrust movement towards NW. The fabric patterns are thus brought in as evidence in the controversy on the age of the Moine schists in favour of a Pre-Caledonian (Pre-Cambrian) age of this group of rocks. A comparison with the Bygdin area will at once annihilate this argument, even with the rejection of the interpretation of similar structures here presented. In the case of the Bygdin area the possibility of a Pre-Caledonian age of the rocks involved (conglomerates in the Valdres Sparagmite) is excluded, nor can there thus be any Pre-Caledonian deformation.

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[1] The same opinion as to the significance of girdles has been expressed by Kvale (1941, p. 197).

The structural analogy between the Moine schists and the rocks here studied is of great interest as an indication of a regularity in structures, including those of petrofabric patterns, resulting from similar tectonical conditions. As sufficient material accumulates, structural petrology may prove of value for the interpretation of tectonical structures, perhaps especially in old zones of deformation.

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To the diagrams.

☐	░	▥
< $\frac{1}{2}$%	$\frac{1}{2}$ - 1%	> 1-2%
▤	▦	■
> 2 3%	> 3-4%	> 4%

All the diagrams are in equal area projection. They have all been rotated into the horizontal plane and are projected on the *upper* hemisphere. The schistosity (or bedding) planes are marked by semicircles and the stretching directions by rings.

D 1—19 are from deformed quartzite boulders in the conglomerate, the localities of the specimens from which the diagrams were prepared are marked on map fig. 1 (p. 15). D 20—22 are from sparagmites, localities at the road 1—2 km south of the area of map fig. 1. See also text pp. 20—21.

E 31 ø 430 etc. is the number of the specimen from which the diagram was prepared, in the collection of the Geol. Survey of Norway. 200 qu.—200 quartz axes, 110 ser.—110 poles of sericite flakes.

D 1.	E 31 ø 430.	206 qu.
D 2.	E 31 ø 432.	250 qu.
D 3.	E 31 ø 433.	250 qu.
D 4.	E 31 ø 436.	240 qu.
D 5.	E 31 ø 252.	230 qu.
D 6.	E 31 ø 252.	110 ser.
D 7.	E 31 ø 257.	210 qu.
D 8.	E 31 ø 434.	260 qu.
D 9.	E 31 ø 290.	210 qu.
D 10.	E 31 ø 290.	120 ser.
D 11.	E 31 ø 271.	260 qu.
D 12.	E 31 ø 435.	300 qu.
D 13.	E 31 ø 256.	235 qu.
D 14.	E 31 ø 256.	100 ser.
D 15.	E 31 ø 428.	210 qu.
D 16.	E 31 ø 428.	120 ser.
D 17.	E 31 ø 426.	230 qu.
D 18.	E 31 ø 438.	240 qu.
D 19.	E 31 ø 427.	210 qu.
D 20.	E 31 ø 243.	300 qu.
D 21.	E 31 ø 241.	300 qu.
D 22.	E 31 ø 242.	220 qu.

D 23. Poles of bedding and schistosity planes (dots) and stretching directions (rings) from the area of map fig. 1. See text p. 17.

D 24. Poles of joint-planes from the area of map fig. 1. See text p. 18.

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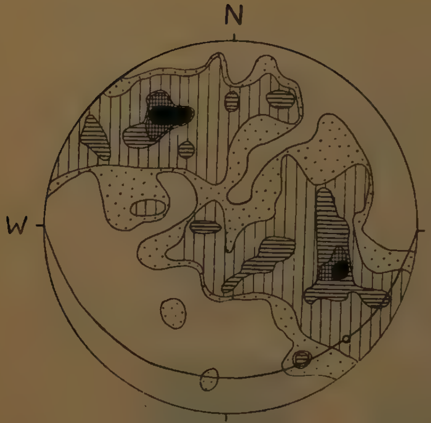

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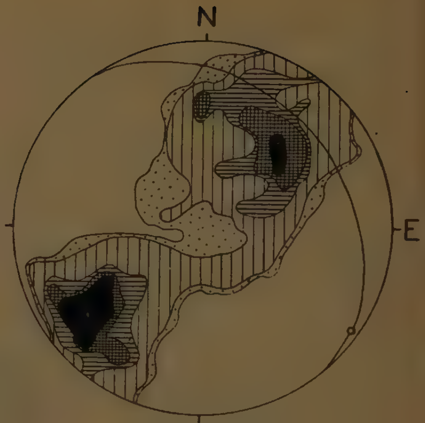

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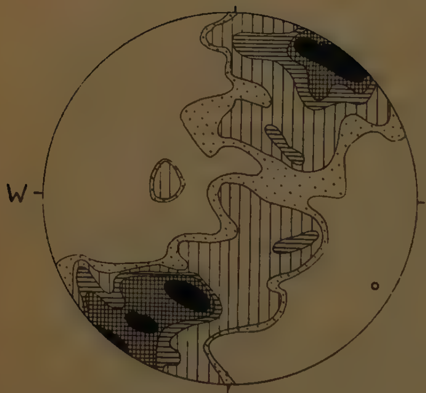

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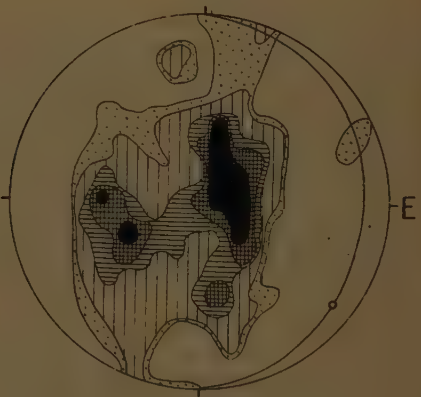

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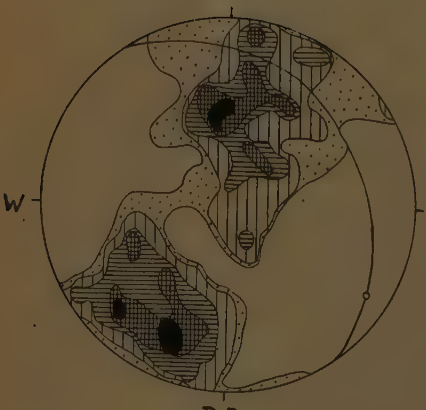

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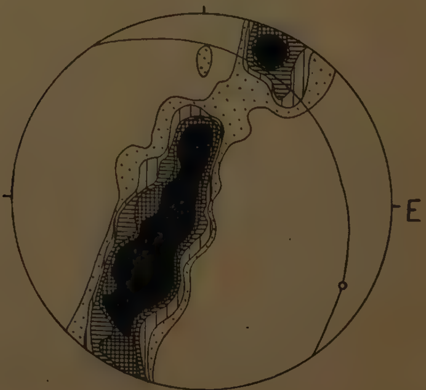

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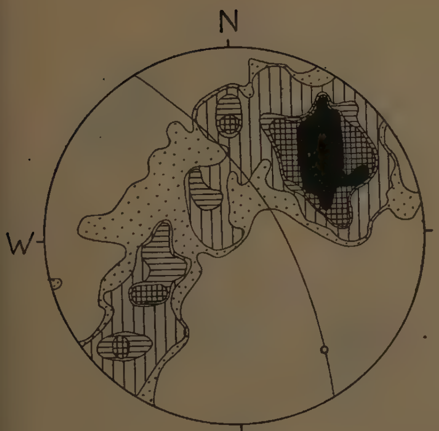

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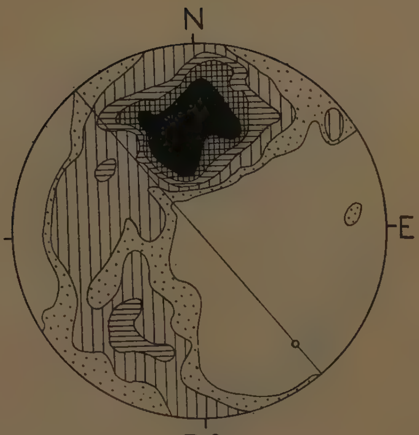

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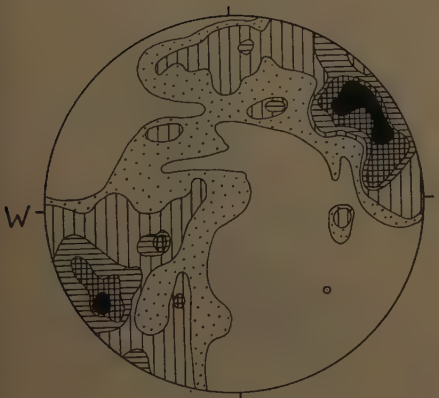

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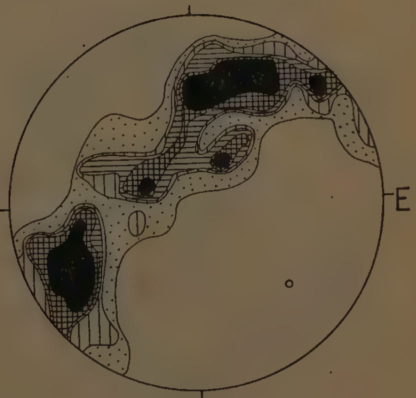

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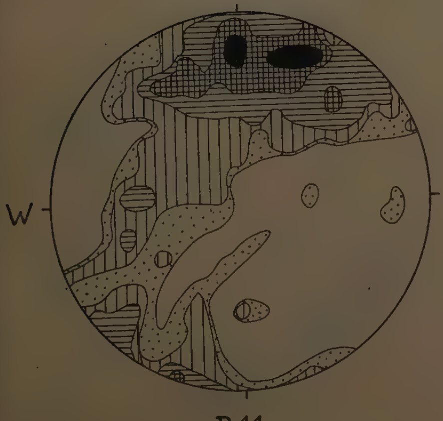

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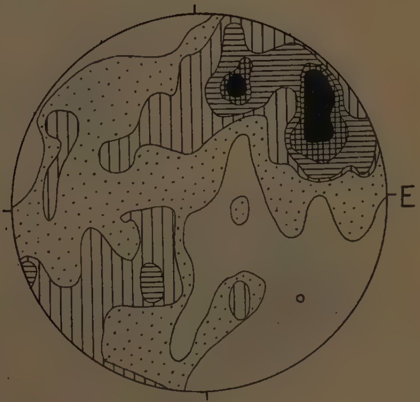

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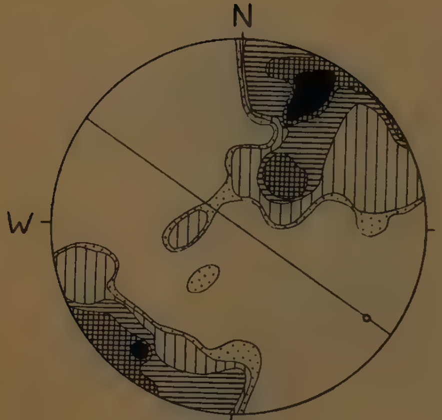

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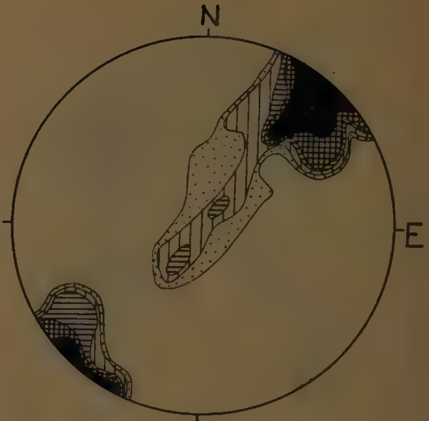

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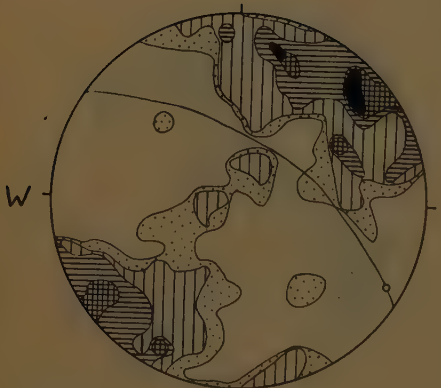

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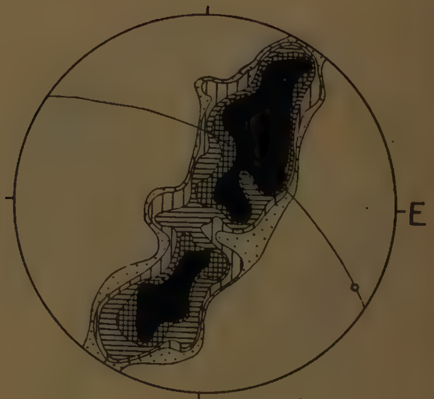

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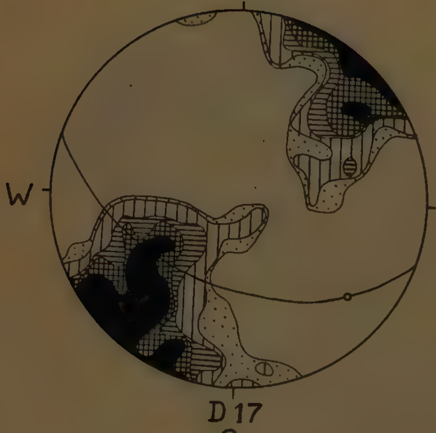

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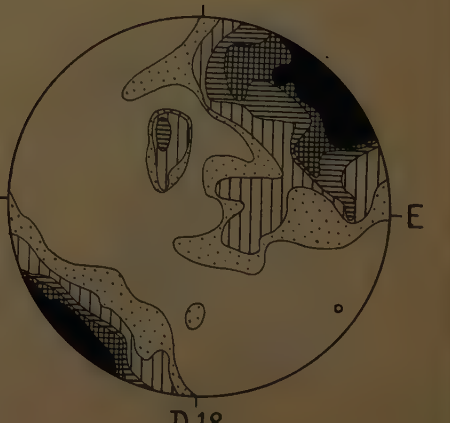

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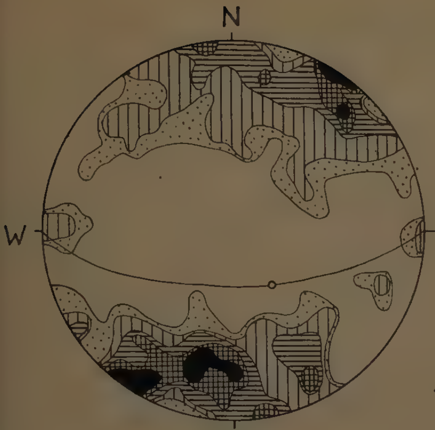

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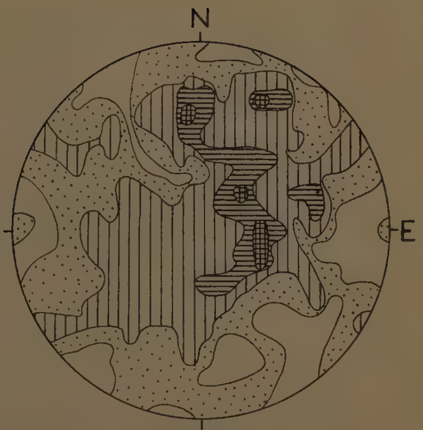

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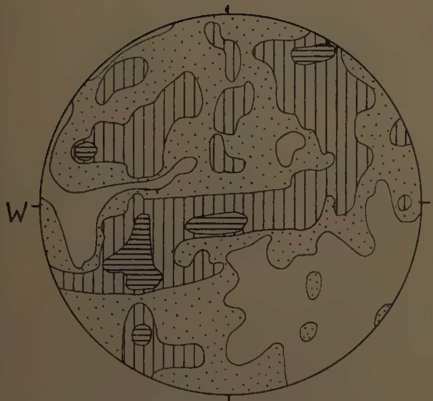

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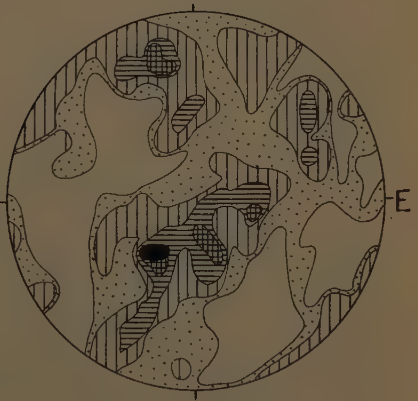

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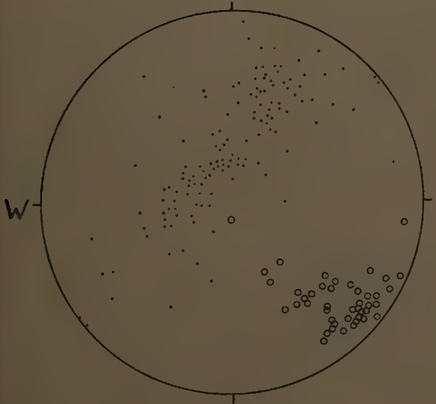

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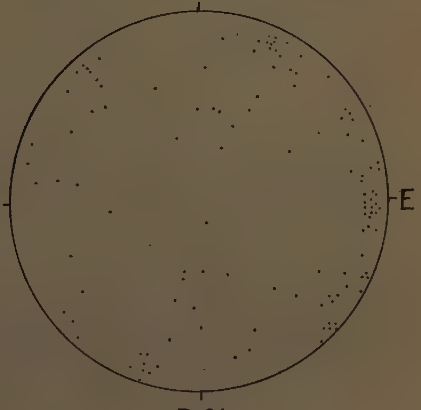

D 24

Ms. mottatt 10. juni 1944.

SKYVNING OG FRIKSJON

AV

ANDERS KVALE

Med 1 fig i teksten. Engl. Summary.

Det er meget fortjenstfuldt av CHR. OFTEDAHL at han i sitt arbeide om sparagmitten og dens skyvning[1] diskuterer den mekaniske side av skyveproblemet. Selv om vi ennå er langt fra løsningen, er det ønskelig at problemene blir klarlagt, og at det blir pekt på vanskelighetene.

Oftedahl forsøker å beregne friksjonen ved skyvning av sparagmittflakene og finner at de krefter som skulle til for å overvinne den måtte være betydelig større enn bergartens „ultimate strength" for brudd. Hvis sparagmittflakene skal ha overført det trykk som frambragte foldningen i Oslofeltets bergarter ved Mjøsa, må friksjonen ha vært like liten som for et jernbanetog.

Dette friksjonsproblemet er tilsynelatende uløselig. Det har ført til at enkelte, særlig amerikanske geologer, som ikke kjenner store overskyvninger fra selvsyn, har vært tilbøyelige til å tvile på deres eksistens. Bevisene for skyvninger i stor målestokk er imidlertid nå så sterke at en ikke kan tvile på dem. Vi kan vel si at de fleste norske geologer nå tror på overskyvninger i fjellkjeden, og det fyldige materiale som Oftedahl legger fram betyr en ytterligere styrkelse av skyveteorien. En kan heller ikke tvile på de friksjonslover som er funnet ved laboratorieforsøk. Feilen i regnestykket må altså ligge i anvendelsen av laboratoriets lover på forholdene i naturen. Der må gjøre seg gjeldende faktorer hvis virkning vi ikke har full oversikt over, og som bringer inn vesentlige korreksjoner i beregningen.

Lignende vanskeligheter møter en også på andre områder av av tektonikken, således ved anvendelse av strainellipsoidet på deformasjoner i bergarter.[2]

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[1] CHRISTOFFER OFTEDAHL: Om sparagmitten og dens skyvning innen kartbladet Øvre Rendal. N. G. U. Nr. 161. Oslo 1943.

[2] D. T. Griggs: The Strain Ellipsoid as a Theory of Rupture. Am. Journ. Sci. 5th. ser. Vol. 30, pp. 121—137.. 1935.

I det følgende skal nevnes enkelte faktorer som vil innvirke på Oftedahls regnestykke.

Han betrakter sparagmitten som en stor sammenhengende plate, som angripes av en kraft i den ene ende og derved skyves over underlaget. Platen er under bevegelsen så stiv at den kan overføre trykk til Oslofeltets lagpakker, som derved blir foldet. Det er nødvendig å forenkle billedet for å kunne gjennomføre beregningen, men spørsmålet er om ikke denne forenkling er litt for stor.

Oftedahl skiller mellom tre former for skyvning:

1. Den differensielle skyvning.
2. Skyvning av enkelte benker.
3. Skyvning av hele lagpakken.

1. *Den differensielle skyvning.* En rekke detaljer i strukturbeskrivelsene viser at sparagmitten har vært utsatt for et betydelig stress, som har ført til oppknusning med rekrystallisasjon og strekning av konglomerater, slik at forholdet mellom deres minste og lengste akse er blitt opptil 1 : 10. Dette viser at påvirkningen har oversteget elastisitetsgrensen betydelig, og at sparagmitten har reagert ved plastisk „flow“. De enkelte prosesser ved en slik „flow“ er i stor utstrekning klarlagt ved arbeider av Sander, Schmidt m. fl. Bergarten har vært i en tilstand hvor den kunne deformeres plastisk, det vil si at den har reagert på stresset ved partialbevegelse uten brudd. En mindre del av bergarten har vært fluid, mineralene har vært påvirket av skjærkrefter som har rotert dem inn i visse stillinger, det har sannsynligvis foregått en glidning langs visse gitterplan og samtidig har det foregått en rekrystallisasjon. Hvor stor friksjonen er i en bergart under slike forhold er ikke godt å si. Sander pleier å sammenligne en slik „tektonisk transport“ med en strømmende elv. Ideen er bra nok, hvis en bare ikke tenker seg at elven inneholder vann, men heller bek, eller et enda seigere stoff.

Et av de spørsmål som ennå ikke er klarlagt, er strekningsretningens forhold til bevegelsesretningen. I Alpene er som regel foldningsakser og strekning loddrett på bevegelsesretningen. Sanders og Schmidts teorier bygger på at dette er det normale forhold, og de kan gi en tilfredsstillende forklaring på dannelsen av denne strekning. Flere faktorer kan spille inn, men viktigst er en rotasjon om en akse loddrett bevegelseretningen, noe i likhet med deig som rulles på et bord. I den kaledoniske fjellkjede opptrer en annen type av strekning, nemlig omkring OSO til SO, altså parallell den skyveretning som alle fjellkjedegeologer på begge sider av Kjølen er enige om. Denne strekning finnes på en rekke steder fra Ryfylke til Torneträsk. I enkelte strøk er den helt dominerende, og ettersom fjellkjedeundersøkelsene skrider framover, får man inntrykk av at den er den viktigste strekningsretning over store strøk. Foldning med akser i denne retning forekommer også, men som regel i liten målestokk. En slik strekning er ikke lett å forklare etter Sanders teorier, særlig ikke når diagrammer av bergartenes mikrostrukturer viser alle de egenskaper som kjennetegner B-tektonitter, dvs. at bevegelsesretningen skulle være loddrett strekningsretningen. Slike diagrammer førte F. C. Phillips[1] til å anta at strekningen parallelt skyveretningen i Nord-Skottland var uten forbindelse med skyvningen, men skyldtes en eldre deformasjon med bevegelsesretning loddrett den senere skyveretning. Muligens er forholdene i Skottland annerledes enn hos oss, men en vilde komme til samme konklusjon i vår fjellkjede hvis en bare skulde holde seg til de gjeldende teorier.

Oftedahls uttalelser s. 54—55 tyder på at strekningen på bladet Øvre Rendal som regel er i retning NV—SO og parallell skyveretningen. Lokalt opptrer linjestruktur, tildels også strekning av konglomerater, etter retningen NO—SV. Oftedahl antyder at den skulle skyldes bevegelse loddrett den vanlige. Det tror jeg en skal være forsiktig med. Det forekommer meg mer sannsynlig at det her er strekningen loddrett bevegelsesretningen som av en eller annen grunn er kommet til uttrykk. Den samme veksling mellom de to hovedretninger er kjent fra en rekke andre steder i fjellkjeden. Om årsaken til denne veksling vet vi foreløbig svært lite, men jeg skal forsøke en forklaring. En viktig betingelse for dannelsen av strekning i bevegelsesretningen er at bevegelsen er laminar, det vil si at det ikke foregår rotasjon om en akse loddrett bevegelsesretningen. Hvor det opptrer strekning loddrett bevegelsesretningen har det ofte foregått en slik rotasjon; finner vi foldningsakser parallelt strekningen, er rotasjonen bevist. En slik rotasjon kan skyldes ujevnheter i underlaget, på samme måte som det i en elv dannes hvirvler der bunnen er steinet. Vi kan derfor tenke oss at hvor skyvningen har gått jevnt uten særlige hindringer, er det blitt strekning i skyveretningen, men hvor underlaget har vært ujevnt, er strekningen blitt loddrett skyveretningen. Hvis en rotasjon ikke kan påvises, må en prøve andre forklaringer. I alle tilfelle kan spørsmålet ikke avgjøres uten inngående undersøkelser.

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[1] F. C. Phillips: A Fabric Study of some Moine Schists and Associated Rocks Quarterly Journal of the Geologic Society Vol. — B pp. 581—616. London 1937.

Dette med strekningsretning og bevegelsesretning er litt av en digresjon, men jeg vilde gjerne få framheve betydningen av at de som arbeider i fjellkjeden legger merke til de retninger for strekning, foldningsakser og andre linjestrukturer som opptrer, og hvorledes de forholder seg til den eller de skyveretninger som en mener å kunne påvise i vedkommende strøk.

I hvilken grad de enkelte mineralkorn i sparagmitten er blitt reorientert under deformasjonen vet vi ikke; men det er grunn til å tro at der hvor strekningen har vært sterkest, vil orienteringen vise bestemte mønster. Jeg holder det sannsynlig at bergartene mange steder vil vise seg å være B-tektonitter med strekningsretningen som B-akse.

For å kunne bedømme størrelsen av den differensielle skyvning beregner Oftedahl på s. 47 gradienten på grunnlag av akseforholdene i et deformert konglomerat. Framgangsmåten er ikke klar, men den verdi han finner, 3, er den samme som Mügge[1] fant som minimumsgradient i en glimmerskifer med granater hvis rotasjon kunde måles.

2. *Benkenes skyvning.* Oftedahl har påvist at de enkelte sparagmittbenker i atskillig utstrekning er blitt skjøvet på smale soner av mylonittskifer hvis sammensetning ligger nær det marine leirs. En slik oppdeling av skyvningen er å vente hvor lag av så forskjellig sammensetning veksler. Selv i ensartet bergart vil en middelssterk deformasjon ofte være konsentrert i enkelte soner. Jeg har således sett konglomerater hvor tilsynelatende udeformerte lag vekslet med lag hvor rullesteinene hadde en betydelig strekning. Finnes det i en bergart lag som byr mindre motstand mot skyvning enn resten av bergarten, så vil bevegelsen for en vesentlig del foregå langs disse lag. Sammenligningen med et smøremiddel er meget god, også i den forstand at smurningen brukes opp under transporten. På Vestlandet er det glimmerskifer som har vært smurning under skyvningene. En finner smale striper av den nær sagt over alt hvor det kan påvises å ha foregått skyvning. På kartblad Bergsdalen er det en slik glimmerskifer som i kartets NO hjørne er opptil 1200 m tykk, men

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[1] O. Mügge: Bewegungen von Porphyroblasten in Phylliten und ihre Meßung. Neues Jahrb. für Min. Geol. und Pal. Beil. Bd. 61, Abt. A, s. 469—510. 1930.

blir smalere etter hvert som vi kommer i strøk med mer intens deformasjon. Til slutt er den bare noen få meter bred og er helt forsvunnet på enkelte strekninger. Det skulde være grunn til å anta at det også under sparagmittbenkenes skyvning har vært slitasje på smurningen, slik at det som nå er mylonittskifer opprinnelig hadde en større utbredelse enn en finner i dag.

Oftedahl skriver s. 45 at mylonittskifrene „kun synes å optre på steder hvor skjærkrefter eller sterkt trykk har virket (eller omvendt)“. Hvis det med parantesen menes at i alle lagserier hvor denne skifer fantes, er bevegelsen helt overveiende knyttet til den, synes jeg at denne formulering er atskillig mer tiltalende.

3. *Skyvning av store flak.* Det ble nevnt ovenfor at en strekning i bevegelsesretningen er vanskelig å forklare. Særlig blir det vanskelig hvis vi tenker oss en plate som skyves framover ved et trykk bakfra. Strekningen vilde bli mer forståelig hvis platen var blitt dradd framover, slik at den ble strukket i bevegelsesretningen. En slik strekkende virkning kunde tenkes hvis det over platen lå en annen plate som beveget seg med større hastighet. Oftedahl diskuterer ikke på hvilket dyp skyvningen har foregått, men nydannelsen av sericitt og rekrystallisasjonen av kvarts viser at temperaturen har vært omkring et par hundre grader. Friksjonen har nok bevirket økning av temperaturen, men hadde skyvningen foregått like opp i dagen vilde vi fått rene mylonitter. Det har altså ligget en bergartserie på i hvert fall noen hundre meter over dem vi nå finner, og disse har etter all sannsynlighet beveget seg med større hastighet. Oftedahl hevder at gradienten synes overalt å være positiv oppover, og det er meget sannsynlig, da motstanden mot bevegelsen blir mindre jo nærmere en kommer mot overflaten. Vi får her analogi med en elv, hvor jo hastigheten er størst i overflaten og minst nær bunnen. I en elv vil de øvre vannlag søke å trekke de undre med seg, og en lignende virkning skulde ikke være utelukket ved tektonisk transport. Denne medslepning alene er ikke nok til å forklare strekningen, men jeg holder det sannsynlig at den er en medvirkende faktor.

Kanskje det viktigste spørsmål i forbindelse med problemet skyvning og friksjon er kreftenes angrepspunkt. Oftedahl regner med et trykk som har angrepet langt opp i nordvest og forplantet seg gjennom sparagmitten helt ned til Oslofeltets kambro-silur. Hans beregninger gir da en friksjon som for et jernbanetog. Hvis vi nå kunde anta at kreftene ikke bare angrep lengst bak i sparagmittflakene, men at de angrep over et større område, om ikke over det hele, så vilde vi kunne regne med friksjonskoeffisienter som kom de sannsynlige verdier meget nærmere. Er det da noen mulighet for en slik virkning av kreftene?

W. Schmidt[1] hevder at større skyvninger ikke kan finne sted uten at det virker massekrefter som angriper i hvert punkt av skyveflaket, og som er store nok til å overvinne friksjonen. Trykket i lengderetningen blir da uten betydning, bevegelsen blir som i en elv, hvor tyngdekraften er den eneste bevegende kraft. Et trykk bakfra vil alene bare kunne framkalle ubetydelig bevegelse, mener han. Av hvilken art disse massekrefter er, diskuterer han ikke.

Schmidts oppfatning kan nok være litt ensidig, men det ser ut til at vi ikke kan komme utenom medvirkningen av slike massekrefter ved store skyvninger. Skal vi drøfte deres natur, kommer vi inn på årsakene til fjellkjedefoldninger, og da er vi med en gang på gyngende grunn. Det ser ut til at konveksjonsstrømninger i magma for tiden har de fleste tilhengere. Disse strømninger fører til tangentiale bevegelser like under den faste jordskorpe, men at de skulde kunne virke som massekrefter på et skyveflak nær jordoverflaten er ikke lett å forestille seg.

Blant de andre foreslåtte krefter som skulde framkalle fjellkjedefoldninger er både centrifugalkraften og tidevannskraften massekrefter; men begge er etter Jeffreys beregninger så små at det er utelukket at de kan ha noen virkning. Det har nok hendt at geofysikernes beregninger har måttet korrigeres; men i dette tilfelle er forskjellene så store at noen nevneverdig virkning av disse krefter synes utelukket.

Et spørsmål av betydning for skyveproblemet er også hvilken stilling skyveplanene hadde under bevegelsen. Hvis de hadde sin helling i skyveretningen, vil også tyngdekraften komme inn som en virksom faktor. Haarmann har bygget opp sin oscillasjonsteori på dette grunnlag. Den har fått meget motbør, men også flere av dens motstandere regner med muligheten av skyvninger nedover f. eks. i Alpene.

En liten overlegning vil vise hvilken rolle tyngdekraften spiller for skyveproblemet. Av figuren sees at når skyveplanets hellingsvinkel er V, blir tyngdekraftens komponent parallelt skyveplanet $P_1 = P \sin. V$. For $V = 17{,}5°$ blir $P_1 = 0{,}3$ P og for $V = 30°$ blir

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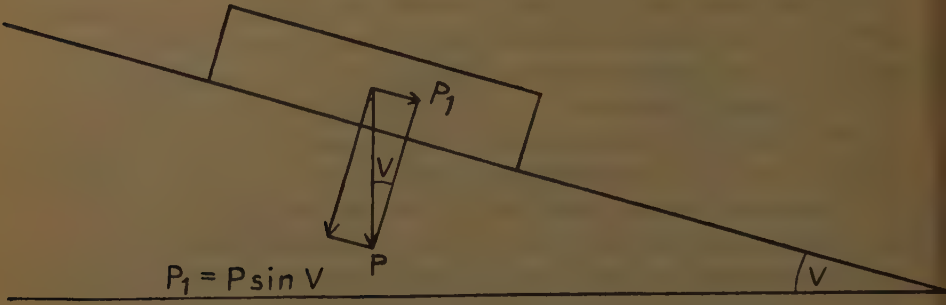
[1] W. Schmidt: Tektonik und Verformungslehre. Berlin 1932, s. 153.



Ved skyvning på et skråplan blir tyngdekraftens komponent parallelt skyveplanet $P_1 = P \sin V$.

$P_1 = 0{,}5$ P, de samme verdier som Oftedahl regner med for friksjonen under sparagmittens skyvning. Ved helningsvinkler på 20—30° skulle altså tyngdekraften kunne overvinne i alle fall det vesentligste av friksjonen mot underlaget.

Er det da noen grunn til å tro at skyveplanet kan helle nedover? En skyvning foregår alltid fra de centrale deler av en fjellkjede mot periferien Etter de vanlige teorier antar en at de centrale deler av fjellkjeden vil presses ned på store dyp og senere heves opp så de ligger høyere enn landet på begge sider. Hvis det foregår skyvninger i denne fase av en orogenese, er det grunn til å tro at i hvert fall enkelte skyveplan kan helle nedover.

Et spørsmål av interesse i denne forbindelse er om skyveflaten fører ut i dagen eller ikke. I siste fall må en regne med motstanden fra de masser som skyves foran dekket. W. Schmidt hevder at større skyvninger bare kan komme i stand hvis det dannes en skyveflate som fører ut i dagen. Han bygger sin oppfatning på teoretiske spekulasjoner, og det vil nok bli vanskelig i det enkelte tilfelle å finne beviser for eller mot hans oppfatning.

Selv om skyvningen i mange tilfelle kan ha foregått nedover, er dette ingen almindelig regel. Det må ansees bevist at den kanskje like ofte har foregått oppover, og tyngdekraftens komponent parallelt skyveflaten har da måttet overvinnes i tillegg til friksjonen.

Under alle omstendigheter må vi ta tyngdekraften med i beregningen, unntatt ved de bevegelser hvor skyveflaten har vært absolutt horisontal.

Spørsmålet om kreftene er ennå på langt nær ikke løst. Og så lenge det er tilfelle, er vi ikke tvunget til å regne med et angrepspunkt i skyveflakets ene ende som eneste mulighet.

Jeg kan ikke sette opp det riktige regnestykke for en skyvning, men jeg har forsøkt å peke på noen av de faktorer som er med og kompliserer beregningen, og som gir oss grunn til å tro at det hverken er noe i veien med naturlovene eller med vår tro på skyvningens eksistens.

SUMMARY

This paper comments briefly on a recent discussion by Chr. Oftedahl of the mechanics of thrusting, in which is concluded that although the thrusting of eo-Cambrian sparagmite in eastern Norway is proved beyond doubt, the coefficient of friction must have been far below any reasonable limit to permit the movement of large masses more than 100 km.

Several factors are mentioned which complicate the calculations by Oftedahl, e. g. the internal friction of a rock that is stretched and recrystallized during the thrusting, and the possible greater pre-tectonic occurrence of argillaceous beds, which served as lubrication and were smeared out during the movement.

The question of where the thrusting forces attack is considered most important. If one postulates a stress acting in one end of a large thrust sheet as the only agent, the problem of friction seems unsolvable. If forces acting upon every point in the thrust sheet ("Massenkräfte" in the sense of W. Schmidt) were also active, the friction presents no difficulty. At the present state of the theories of mountain-building the action of such forces is by no means precluded. As an example is mentioned the force of gravity, which enters the calculation as an accelerating or retarding force except when the thrust plane is horizontal.

It is also pointed out that in the Caledonides of Norway and Sweden the thrust rocks are very commonly stretched parallel to the direction of movement, while in the Alps the stretching is as a rule normal to this direction. The importance of careful observations on this problem by all geologists working in the Caledonian mountains is stressed.

Trykt 14. februar 1945.

DIREKTØR SVERRE BLEKUM

Minnetale holdt i Norsk geologisk forening 4. mai 1944

AV

ARNE BUGGE

SVERRE BLEKUM ble født på Lindesnes fyr 13. februar 1886. Han var sønn av fyrforvalter Olaus Blekum.

Etter i 1912 å ha fullendt eksamen som bergkandidat, bodde han et par år i Oslo.

Sommeren 1914 var han med på F. Hiorts ekspedition til Spitsbergen.

1915 ble han ansatt som kjemiker ved Knaben Mines i Fjotland, 1916 ble han vaskeri-ingeniør og senere driftsbestyrer av Ørnehommen grube som ligger nær ved den nå i drift-værende Knaben II grube.

Sammen med professor Jakob Schetelig, direktør S. Th. Sverre og advokat H. Jentoft opptok han arbeidet med å samle under en ledelse alle de betydelige skjærp på Knabenheia. Dette arbeide resulterte 12. august 1918 i dannelsen av aktieselskapet Knaben Molybdængruber. Blekum var helt fra starten teknisk direktør i dette selskap inntil han for 2 år siden måtte trekke seg tilbake. Kun i de 2 nedgangsår 1921—22 lå driften helt nede og han bodde da atter i Oslo.

Blekum hadde den for en bergingeniør sjeldne opplevelse å se, at de skjærp — han begynte å legge sitt arbeide i, — under hans tekniske ledelse — utvikledes til en stor grube som da krigen begynte ga malm til en virkelig stor-bedrift.

Han hadde alltid stor interesse for den geologiske undersøkelse av den grube han bestyrte og han søkte allerede ved driftens begynnelse samarbeide med geologer, først med professor Jakob Schetelig, og etter hans død har jeg utført det geologiske arbeide.

I Tidsskrift for Kemi og Bergv. Nr. 9 1932 har Blekum gitt en beskrivelse av A/S Knaben Molybdængruber og kommer der også inn på forekomstens geologi.

Som en helt personlig iakttagelse bør det nevnes at Blekum er den første som har erkjent betydningen av det såkalte gangfjell. Dette gangfjell eller den malmførende bergart gir han følgende karakteristikk:

1. Bergarten er tydelig stripet med utpreget stresstruktur.
2. Fattig på mørke mineraler.
3. Rik på kvarts (både primær og sekundær).
4. Fører konstant impregnasjon av sulfider: molybdenglans, kobberkis, magnetkis og svovelkis.

Erkjennelsen av dette „gangfjell“ som den malmførende bergart, har vært av stor betydning for skjærpearbeidet.

Senere kartlegging har vist at gangfjellet er den mineraliserte granit som forekommer langs sidene av og ved utkilingen av de yngre aplitganger og er dannet som et resultat av den hydrotermale og pneumatolytiske virksomhet som har fulgt etter aplitens framtrengen.

Når vi i denne forening minnes direktør Blekums faglige innsats, bør vi særlig huske at han har vært blant de første grubedirektører i vort land som knyttet en geolog til bedriften som fast medarbeider.

Uten et nær samarbeide mellom det praktiske livs menn og geologene kan geologisk forskning ikke få den plass i vårt samfund som den bør ha. Vi bør derfor minnes ham som en av de framste blant dem i vårt land som har hevet geologien opp til et nyttefag.

Personlig vil alle som har truffet Sverre Blekum minnes ham for hans elskverdige og vennesæle vesen. Han var høyt aktet blant sine kolleger og i den bygd hvor han hørte hjemme og hans gjestfrie hjem står som et lysende minne for mange som etter den lange kjøretur opp gjennom Kvinsedal har nådd fram til Knaben Gruber.

Trykt 16. februar 1945.

Ms. rec. sept. 1944.

PETROLOGICAL SIGNIFICANCE OF SUB-SOLIDUS PHASE TRANSITIONS IN MIXED CRYSTALS

BY

H. Ramberg

With 15 figures in the text.

Abstract. The phase transitions of mixed crystals at sub-solidus conditions are discussed from a theoretical physico-chemical point of view.

Two different types of sub-solidus transitions are found to exist in rock-forming solid solutions. (1). The eutectic type in which only the one component of a mixed crystal series changes into another modification, decomposes, or reacts with other minerals. Under these conditions the transformation or reaction point (or rather—line) in the P,T-diagram varies with varying composition of the solid solution in an analogous manner as the melting or saturation point in eutectic melts or liquid solutions. And (2.) the isodimorphic type by which both end members of a two component mixed crystal change to other modifications, decompose, or react with other minerals. In that case the sub-solidus transition equilibrium is analogous to the melting equilibrium of mixed crystals.

12 examples from rock-forming silicates are discussed broadly. These examples show that the transformations in mixed crystal series can be used as suitable continuous geological P,T-registrators. They further show that there exists gradual rather than discontinuous transitions between the several critical or typomorphic parageneses of the different mineral facies of rocks.

According to papers dealing with metamorphism the chemical composition of some mixed crystals among the common minerals varies with varying degree of metamorphism.

In this connection we may ask if the varying composition of the mixed crystals is due only to differences in the chemical composition of the rocks under consideration, or if there also exists a law of relation between the metamorphic grade and the composition of the mixed crystals. If the latter be the case, the petrologist has very suitable geological temperature and pressure indicators in this inter-relation between the composition of the mixed crystals and the P, T-conditions. The great advantages of such P,T-indicators is the possibility of a continuous registration of temperature and pressure during the metamorphism.

In fact, the petrographic experience suggests a regular relation between the metamorphic grade and the composition of some mixed crystals among rock-forming minerals. It may be well-known, for instance, how the composition of the garnet group varies from Fe-rich almandine in low grade schists to Mg-rich pyrope in high grade eclogites and granulites.[1] Additional examples are the variations in the composition of hornblende and plagioclase with varying degree of metamorphism. The hornblende usually grows richer in Al and $Fe^{\cdot\cdot\cdot}$ as the P,T-conditions increase. The plagioclase is enriched in the anorthite-molecule with increasing metamorphic grade. The varying $Fe^{\cdot\cdot}/Mg$ ratio in biotite is often interpreted as a result of variations in the metamorphic conditions whereunder the biotites are formed. But, since $Fe^{\cdot\cdot}$-rich biotites sometimes are found in low grade rocks, sometimes in high grade rocks, we cannot unconditionally take a regular relation between the $Fe^{\cdot\cdot}/Mg$ ratio in biotite and the P,T-conditions for granted.[2] This variation in the biotite is probably more governed by the chemical environment than by the physical conditions.

To be able to employ the important possibilities which here seem to stand at the disposal of the geologist in determining the P,T-conditions during the reactions in the earth's crust, we have to investigate accurately the physico-chemical laws of the mixed crystals. We have to find the relation between the average chemical composition of the mineral assemblage under consideration, the composition of the mixed crystals in the assemblage, and the P,T-conditions.

Before discussing the physico-chemistry of the mixed crystals, we shall refresh our memory concerning the elementary theory of the chemical stability of the compounds.

Independent on the way of reaction, the thermodynamically stable parageneses are determined by the average chemical composition of the system under consideration, and the gas tension of the different compounds which may be formed under the given P,T-conditions. The stable association consists of the minerals giving minimum partial and also total gas tension.

Since the gas tension of the minerals varies with temperature and pressure, different parageneses may give minimal partial gas tension by varying P and T, and constant average composition.

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[1] P. Eskola in Barth, Correns, Eskola: Die Entstehung der Gesteine. Berlin, p. 364, 1939.

[2] T. F. W. Barth: Bull. Geol Soc. Am. 47, p. 782—784, 1936.

Different assemblages may accordingly be stable at different temperature and pressure, but identical average chemical composition. On this fundamental physico-chemical law the facies classification of metamorphic rocks is based.[1]

According to the said connection between the gas tension and the stability of compounds, the just discussed relation between the composition of mixed crystals and the P,T-conditions whereunder they are formed, indicates a dependence between composition and partial gas tension of mixed crystals. Because of the lack of experiments we have no empirical evidence of the variation of the partial gas tension or of the solubility (which depends on the gas tension itself) with the composition in rock-forming solid solutions. Indirectly some of the melting diagrams of the minerals give important information on the stability and gas tension in the sub-solidus phase. But the most important knowledge of gas tension, stability, and, accordingly, the reaction of solid solutions in common silicates, is due to the analogy conclusion we can draw from the theoretical and experimental metallography.[2]

The relation between partial gas tension and the composition of mixed crystals is fundamental for the understanding of their reactions and stability. Mixed crystals, which also have been called solid solutions, are in many ways similar to liquid solutions. The well-known osmotic laws are applicable both in the liquid solutions and in the solid solutions of the mixed crystals.

For convenience we shall only discuss two-component systems. Then we have: *The partial gas tension of one component in a mixed crystal increases gradually with increasing molecular concentration of the same component in the solid solution.*

In real crystals there seldom or never exists a direct proportionality between the partial gas tension and the molecular concentration in the solid phase. But in so-called ideal mixed crystals the partial gas tension is directly proportional to the concentration, and the partial gas tension may be expressed by the well known Raoult's equation:

$$P_{Am} = P_A \cdot \frac{m}{m+n} \quad \ldots\ldots\ldots \quad (1)$$

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[1] P. Eskola: Norsk geol. tidsskr. Bd. *6*, 1921.

[2] U. Dehlinger: Chemische Physik der Metalle und Legierungen, Leipzig, 1939. O. Kubaschewski: Zeitschrift für Elektrochemie *48*, p. 559 and 646, 1942.

Fig. 1. The partial gas tension of a solution AB in the case of complete miscibility. On ideal osmotic solubility the partial gas tensions follow the solid lines A—P_B and B—P_A respectively. With real solutions the gas tensions lie above or beneath these two lines.

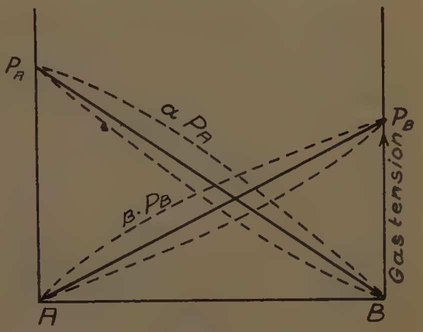

Fig. 1.

where P_A is the gas tension of the pure component A; P_{Am} is the partial gas tension of the composition $A_m \cdot B_n$; m = mol. of A, and n = mol. of B in the solution. (Fig. 1.)

In such an ideal solid solution the chemical affinity between two unequal molecules is identical with the affinity between to equal molecules. Coincidently there must be no distortion in the lattice during the mixing. (This ideal case may perhaps occur in "mixed crystals" containing isotopes: $NaCl^{85}$. $NaCl^{87}$.)

If the unit cells of the two components are more or less different, and the affinity between two unequal molecules is not greater than between two equal molecules, then there is no longer ideal osmotic miscibility between the components, and the curves of partial gas tension lie above the straight lines of fig. 1. If, on the contrary, the affinity between two unequal molecules is greater than the affinity between two equal molecules, and the deformation of the lattice during the mixing is negligible, then the partial gas tension curves may lie beneath the straight lines of fig. 1.

Complete miscibility between two compounds depends on the conditions that the lattice distortion and the repulsion between two unequal molecules do not exceed a given maximum critical value. At these maximum deformation and repulsion values the partial gas tension of a component in a solid solution of definite composition equals the gas tension of the same pure component, or of a solution with a greater concentration of the considered component. On transgressing the mentioned maximum deformations of the lattice, an exsolution of the mixed crystals must take place since the partial gas tension of a component in the solution exceeds the gas tension of the same pure component or of a solution with greater concentration of the same component. (Fig. 2.)

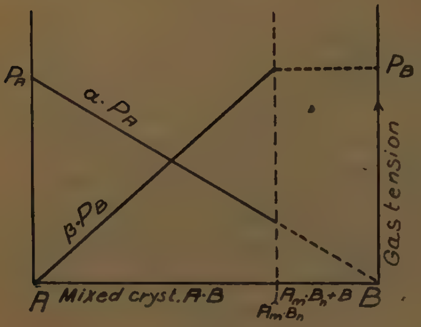

Fig. 2.

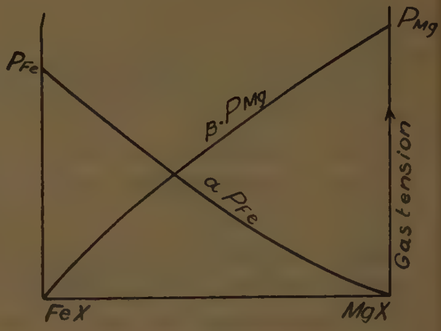

Fig. 3.

Fig. 2. The partial gas tensions of a mixed crystal, AB, in the case of incomplete miscibility. The B-richest end member of the solution ($A_m B_n$) has a B-gas tension which equals the gas tension of the pure B-end member.

Fig. 3. The partial Mg- and Fe-gas tensions of a completely miscible Fe—Mg silicate. (X equals a part of the compounds which is the same for both end members.)

The two-component mixed crystals of metallic alloys usually are of the composition: $A_m B_n$ where A and B are metallic elements. Here it is natural to speak about the partial gas tensions of the components A and B since they are pure elements. In the rock-forming minerals on the other hand, the components which build up the mixed crystals are commonly complicated chemical compounds. Here it is more expedient to consider the partial gas tensions of the substitutable ismorphous elements. (The partial gas tensions of Na and Ca in plagioclase, and of Mg and Fe in hornblende, for instance.)

We know little or nothing about the composition and tension of the gas phase of silicates. Probably the silicate molecules in the gas are more or less dissociated; elementary atoms also must be present. We may for instance speak about the partial oxygen tension of a oxyde.[1]

According to the preceding discussion it is obvious that also the partial gas tension of an ismorphous element in a silicate increases with increasing molecular concentratton in the crystal of the component containing the considered element. In a perfectly miscible Mg—Fe silicate for instance, the partial gas tensions of Mg and of Fe principally vary with the composition as seen in fig. 3.

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[1] G. Grube and M. Flad: Zeitschr. Elektrochemie 45, p. 838, 1939.

After this discussion we shall consider some of the reactions in mixed crystals which are of petrological importance.

I. EUTECTIC TYPE

Eutectic type with modification changes.

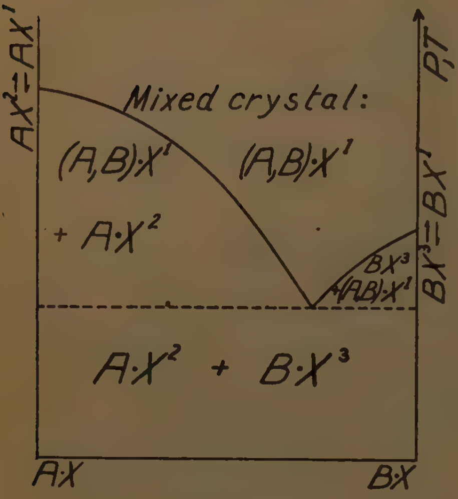
The one component, AX, in a mixed crystal (A,B) X^1 exists in two different modifications: AX^1 and AX^2. Only AX^1 is isomorphously miscible with BX^1. The component BX exists only in the modification BX^1 which is stable all over the stability field of both AX^1 and AX^2. (X be the part ot the compounds which is equal in both the pure components.)

It is important to keep in mind that never the whole amount of the component AX of the mixed crystal (A,B) X^1 is able to separate as the P,T-conditions cross the border-line from modification AX^1 into AX^2.

If the modification AX^2 is the stable one, then the gas tension of AX^2 is lower than that of AX^1. But according to the previously discussed theory of mixed crystals the partial gas tension of A in (A,B) X^1 increases with increasing concentration of the AX-component from zero to a value which maximally equals the A-gas tension of pure AX^1. Consequently, in spite of the gas tension of AX^1 being higher than that of AX^2, there must always exist a definite mixture, (A_nB_m) X^1, the partial A-tension of which equals the partial A-tension of the pure modification AX^2. It is obvious that a mixed crystal of exactly the mentioned composition must exist in stable equilibrium with the pure modification AX^2.

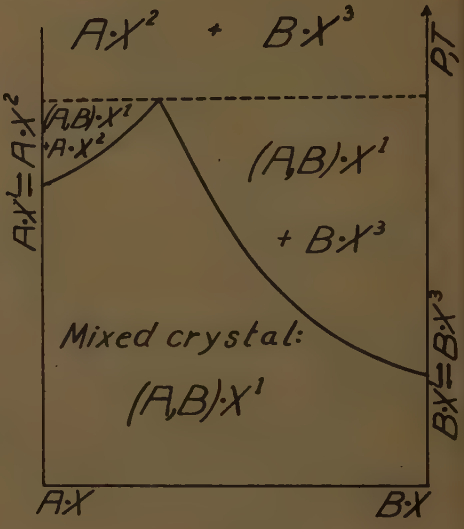
The gas tension of both AX^1 and AX^2 varies with the P,T-conditions. The composition of the mixed crystal the A-tension of which equals the A-tension of the modification AX^2 must therefore vary with varying temperature and pressure.

If AX^1 is the high temperature modification, then AX^2 is in stable equilibrium with gradually A-poorer solid solutions as the temperature decreases below the transformation point: $AX^1 \rightleftarrows AX^2$. With AX^2 as the high temperature modification the same takes place at rising temperatures above the transformation point.



$AX^2 \rightleftharpoons AX^1$
Mixed crystal:
$(A,B)X^1$
$(A,B)X^1$
$+ A{\cdot}X^2$
BX^3
$(A,B)X^1$
$A{\cdot}X^2 + B{\cdot}X^3$
$BX^3 \rightleftharpoons BX^1$
P,T
A·X
B·X

Fig. 4.



$A{\cdot}X^2 + B{\cdot}X^3$
P,T
$A{\cdot}X^1 \rightleftharpoons A{\cdot}X^2$
$(A,B){\cdot}X^1$
$+A{\cdot}X^2$
$(A,B){\cdot}X^1$
$+ B{\cdot}X^3$
Mixed crystal:
$(A,B){\cdot}X^1$
$B{\cdot}X^1 \rightleftharpoons B{\cdot}X^3$
A·X
B·X

Fig. 5.

Fig. 4. Eutectic sub-solidus equlibrium diagram. The mixed crystal $(A, B) X^1$ is the high temperature or pressure modification.

Fig. 5. Eutectic sub-solidus equilibrium diagram. The mixed crystal is the low temperature or pressure modification.

At constant temperature similar conditions exist under varying pressure according as AX^1 is the high or low pressure modification.

The transformation temperature and heat being given, the following equation exists between temperature and composition of the mixed crystal which is in stable equilibrium with the pure AX^2 modification:[1]

$$u\left(1-\frac{T}{T_u}\right)+K_2\,\alpha_2^2+RT\ln(1-\alpha_2)=0 \;\ldots\ldots \qquad (2)$$

u = molar heat of transition
T_u = temperature of transition
α_2 = mol. percent of AX in $(A, B)\,X^1$
K_2 = characteristic constant for the crystal.

The principal equilibrium equation of such phase transition may be stated:

$$(A_mB_n)\,X^1_{(m+n)} \rightleftarrows (A_{m-a}B_n)\,X^1_{(m-a+n)} + a\,AX^2 \;\ldots\ldots \qquad (3)$$

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[1] U. Dehlinger, loc. cit. p. 24.

In the preceding pages it was assumed that the compound BX^1 was stable over the whole stability field of both AX^1 and AX^2. More commonly BX also occurs in two or more modifications, say BX^1 and BX^8. If BX^8 is not miscible with AX^2, then BX^8 exists in equilibrium with $(A, B)\,X^1$ in a similar manner as does AX^2.

These conditions of stability in solid solutions are accordingly analogous to the equilibrium conditions in eutectic melts and liquid solutions. The mixed crystal represents the melt, and the immiscible crystals AX^2 and BX^8 represent the solid precipitates of the eutectic melt or liquid solution. Fig. 4 and fig. 5 are the sub-solidus equilibrium diagrams of eutectic types. In fig. 4 the mixed crystal is the high temperature (pressure) modification, in fig. 5 the immiscible modifications are the high temperature (pressure) phases.

Examples 1 and 2 (p. 52, 53) are of this eutectic type with modifications changes.

Eutectic type with decomposition.

The previously discussed eutectic equilibria must also exist if the component AX in the solid solution $(A, B)\,X^1$ does not occur in two modifications, but decomposes into two or more minerals, none of wich is able to substitute B for A in their lattices.

In this case we have the following equilibrium equation:

$$(A_m B_n)\,X^1{}_{(m+n)} \rightleftarrows (A_{m-a} B_n)\,X^1{}_{(m-a+n)} + b\,AY + c\,AZ \;.\;.\;.\;.\;.\;. \quad (4)$$

where: $b+c=a$; and B is not ismorphous with A in A.Y and A.Z.

Example 3 (p. 54) is of this type.

Eutectic type with reaction products.

As a third possibility of the eutectic equilibria the component AX of the mixture $(A, B)\,X^1$ is assumed to react with a second mineral forming a third mineral. The thus formed reaction-mineral be unable to accept B for A in its lattice. In that case also the P,T conditions effect the equilibrium between a mixed crystals of definite composition and the reaction product. The equilibrium can bee stated as follows:

$$(A_m B_n)\,X^1{}_{(m+n)} + a \cdot CY \rightleftarrows (A_{m-a} B_n)\,X^1{}_{(m-a+n)} + a\,A \cdot C \cdot XY \;.\;.\;.\;. \quad (5)$$

where B cannot bee substituted for A in the compound A. C. X. Y.

Example 5 (p. 58) represents this type of eutectic equilibrium.

II. ISODIMORPHIC TYPE

Isodimorphic type with modification changes.

In the previously discussed eutectic types of sub-solidus equilibria the modification AX^2 which was in stable equilibrium with the mixed crystal $(A\,B)\,X^1$ was unable to accomodate any B-ions in its lattice.

In the general case, however, A and B are mutually exchangeable in all crystal phases.

The simplest case of this equilibrium type is an ideal isodimorphic series i. e. two isodimorphic series $(A, B)\,X^1$ and $(A, B)\,X^2$ the pure components of which are dimorphous modifications of the compounds AX and BX.

Usually the transition points: $AX^1 \rightleftarrows AX^2$ and $BX^1 \rightleftarrows BX^2$ do not coincide, and the transition point for intermediate compositions must therefore vary with varying composition. By considering the partial gas tensions of the solid solutions $(A, B)\,X^1$ and $(A, B)\,X^2$ in a similar manner as was done for the eutectic type p. 47, we arrive at an equilibrium diagram as shown in fig. 6.

At every temperature between the transition temperatures of the pure end components, the two mixed crystals $(A, B)\,X^1$ and $(A, B)\,X^2$ which occur in stable equilibrium, are of different composition. We shall not go through the whole development of the diagram by means of the partial gas tension of the two mixed crystals, we shall only keep in mind that for the two crystals which are in equilibrium, the partial A-tensions of both crystals must be equal at the same time as the partial B-tensions of both crystals are equal.

Since the relation between partial gas tensions and composition are analoguous for mixed crystals and mixed liquids, the equilibrium diagram: mixed crystal$^1 \rightleftarrows$mixed crystal^2 equals the melting diagram of mixed crystals.

For the isodimorphic transition we can use the equation:

$$(A_m B_n)\,X^1_{(m+n)} \rightleftarrows (A_{m-a} B_{n-b})\,X^1_{(m+n-a-b)} + (A_a B_b)\,X^2_{(a+b)} \;.\;.\;.\;.\;.\;. \quad (6)$$

which indicates that a mixture $(A_m B_n)\,X^1_{m+n}$ splits up into two phases of different A/B relations.

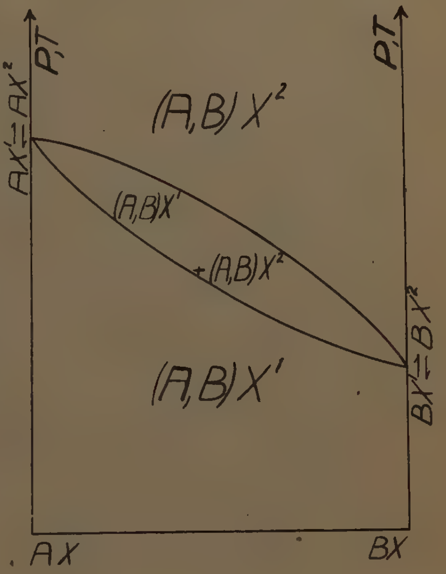

Fig. 6. The principal equilibrium diagram of the sub-solidus transition between isodimorphic series.

Isodimorphic type with decomposition or reaction.

The isodimorphic transition is only a phase transition between two different *modifications* of the end members. Analogous equilibrium must also exist if both the end compounds split up into two or more minerals, or react with other minerals. In these cases the reactions of the end members are:

$$\left.\begin{aligned} AX^1 &\rightleftarrows AY + Z \\ BX^1 &\rightleftarrows BY + Z \end{aligned}\right\} \ldots\ldots \qquad (7)$$

or

$$\left.\begin{aligned} AX^1 + Y &\rightleftarrows AZ + Q \\ BX^1 + Y &\rightleftarrows BZ + Q \end{aligned}\right\} \ldots\ldots \qquad (8)$$

where A and B are substitutable in (A, B) X^1 and (A, B) Y or (A, B) Z recpectively.

Examples 6—12 (p. 60—70) are of these different varities of the isodimorphic types.

III. EXAMPLES

Example 1. The epidote—plagioclase equilibrium.

The equilibrium which exists between plagioclase and epidote (or better zoisite) in low and medium grade rocks is approximately of the discussed eutectic type (p. 47).[1] Epidote or zoisite may be looked upon as the low temperature modification of the anorthite-compound. Consequently there must exist an equilibrium between epidote and a plagioclase mixed crystal of gradually anorthite-poorer composition as the temperature decreases below the transition point: anorthite$\rightleftarrows$zoisite. (Fig. 7.) The molar volumes of anorthite and zoisite indicate that the latter is the high pressure modification of the two minerals, hence the plagioclase which is in stable equilibrium with zoisite must grow gradually richer in albite as the pressure increases above the transition pressure: anorthite$\rightleftarrows$zoisite by constant temperature. Temperature and pressure act against each other on the zoisite—plagioclase equilibrium. Under regional metamorphic conditions the influence of the temperature upon the equilibrium exceeds the influence of the pressure, and accordingly the plagioclase successively must grow richer in anorthite by increasing metamorphic grade. The equation can be given as:

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[1] T. Strand: Norges geol. unders. Nr. 159 p. 43—47. 1943.
I. Rosenqvist: Norsk geologisk tidsskr. *22*, p. 107—202 1944.
H. Ramberg: Norsk geologisk tidsskr. *23*, p. 120—135. 1944.

$$\underset{\text{basic plag.}}{\overset{\text{High T.}}{An_m \cdot Ab_n}} \rightleftarrows \underset{\text{acid plag.}}{\overset{\text{Low T.}}{An_{(m-x)} \cdot Ab_n}} + y \text{ zoisite} \ldots\ldots \tag{9}$$

But because of the small difference in the chemical composition of anorthite and zoisite the equilibrium: plagioclase$\rightleftarrows$zoisite is not only dependent on the P,T-conditions. The average composition of the paragenesis plays a role. In supersilified and H_2O-saturated rocks only the ratio Ca/Al effects the equilibrium at constant P and T. Since the Ca/Al ratio is greater in zoisite than in anorthite, Ca-rich environments in the pore liquid will displace the equilibrium towards zoisite and albitic plagioclase. Al-rich environments, on the other hand, will displace it towards anorthtitic plagioclase.

Petrographic experience seems to prove that the difference in the Ca/Al ratios of the common minerals: garnet, hornblende, biotite and diopsidic pyroxene has no important effect on the equilibrium plagioclase$\rightleftarrows$zoisite. That is to say that the composition of the plagioclase which is in stable equilibrium with zoisite in common rocks is principally determined by the P,T conditions.

But in rocks carrying calcite the equilibrium is displaced towards almost pure albite and zoisite also at high temperature[1] (reaction skarn rocks) because of the high CaO-concentration in the pore liquid. Whether or not the Al-rich environments in cyanite-carrying rocks move the equilibrium towards basic plagioclase, I have not been able to find out.

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[1] H. Ramberg, Loc. cit. p. 118—131.

Fig. 7. Sub-solidus equilibrium diagram of plagioclase and epidote. The ratio $(Ca, Na_2)/Al$ of the paragenesis which the diagram represents lies between ½ and ⅔. Quartz and water can be present without influencing the transition line: epidote ⇄ plagioclase. I, II, III and IV represent the green schist facies, the epiote amphibolite facies, the amphibolite facies and the granulite facies respectively. At point *x* epidote (zoisite) exists together with a plagioclase of the composition *y*. As the metamorphic grade increases from *x* to *z* the composition of the plagioclase follows the line *yz*, and at point *z* epidote disappears. Concerning the borderlines between the different facies see p. 70—73.

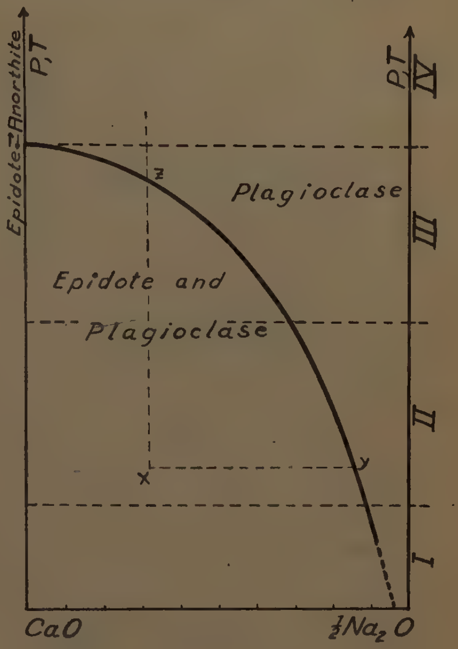

Fig. 7.

Example 2. The talc—anthophyllite equilibrium.

Another illustration of the equilibrium of the eutectic type between minerals, exists in the association: talc-anthophyllite (the name anthophyllite in this connection means the total mixed series between $Fe^{\cdot\cdot}$ and Mg-silicates).

In anthophyllite, $Fe^{\cdot\cdot}$ and Mg ions are mutually substitutable to a high degree. In the lattice of talc only about 2 percent of the MgO may be substituted for by FeO. As all petrographic experiences show, the antophyllite is stable at higher temperatures than talc. Talc may be looked upon as the low temperature modification of pure Mg-anthophyllite. Accordingly there exists an equilibrium between talc and anthophyllite of successively Fe-richer composition as the temperature decreases below the transition-point: talc ⇄ Mg-anthophyllite:

$$\underset{\text{high T}}{\text{Mg-rich anthophyllite} + \text{quartz} + H_2O} \rightleftarrows \underset{\text{low T}}{\text{Mg-poor anthophyllite} + \text{talc}} \qquad (10)$$

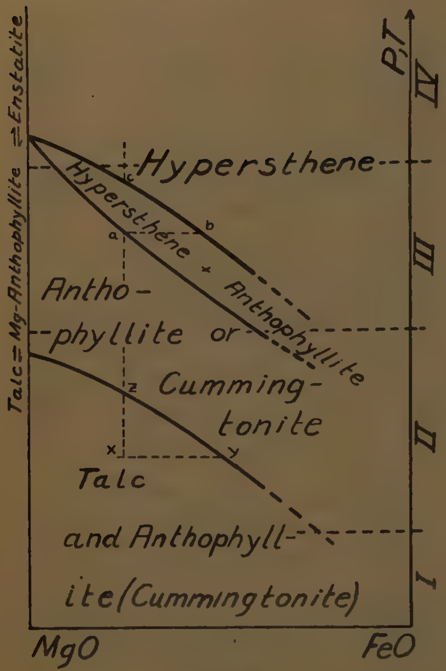
Talc ⇌ Mg-Anthophyllite ⇌ Enstatite
P,T
IV
III
II
I
Hypersthene
c
b
a
Hypersthene
x
Anthophyllite
Antho-phyllite or Cummingtonite
z
x
y
Talc and Anthophyllite(Cummingtonite)
MgO
FeO

Fig. 8.

Fig. 8. Equilibrium diagram of the Fe-Mg silicates at sub-solid state. (Ouartz and water may exist besides the other minerals in the figure.) The lowest part of the diagram shows the talc $\rightleftarrows$ anthophyllite equilibrium of eutectic type. The upper part represents the isodimorphic type of equilibrium between orthopyroxene and cummingtonite or anthophylite. I, II, III, and IV represent the green schist, the epidote amphibolite, the amphibolite and the granulite facies respectively. At point *X*, talc exists together with an anthophyllite of composition (Fe/Mg ratio) *Y*. By increasing metamorphic grade the Fe/Mg-ratio of the anthophyllite which is in equilibrium with talc, follows the line *YZ*. By passing the point *Z* the talc disappears, and only anthophyllite exists up to the point *a* where hypersthene of composition *b* appears. First above point *c* the anthophyllite disappears.

Because of the difference in the Mg/Si and Mg/H_2O ratios in talc and Mg-anthophyllite, the above equilibrium is only valid for supersilified and H_2O-saturated rocks. (Fig. 8.)

The association: talc-anthophyllite is not very common, and the equilibrium between the two minerals is thus of no great importance as a geological P,T indicator. In the study of metamorphic facies in provinces consisting of greatly diversified rocks it is, on the other hand, also important to include critical parageneses of rare rocks.

Example 3. The talc, calcite—actinolite equilibrium.

The paragenesis: actinolite, calcite, and talc is a good example of the eutectic sub-solidus transition with decomposition. (Actinolite means the total mixed crystal series between Fe and Mg-silicate.)

The pure Mg-component of the mixed crystal actinolite — the tremolite — decomposes at a relatively low temperature into talc and calcite according to the equation:

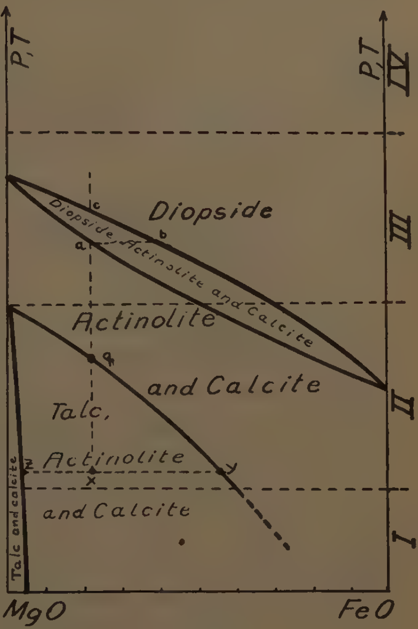

Fig. 9. Equilibrium diagram of the (Fe,Mg)-lime silicates in sub-solidus state. The minerals in the diagram are saturated in silica, water and lime so that quartz, water and calcite may be present besides the other minerals without displacing the equilibrium. The ratio Ca/(Fe,Mg) of the parageneses which the diagram represents is equal to, or greater than unity. At point *X* talc and calcite are in equilibrium with an actinolite of composition *Y*. (The talc may have the Fe/Mg-ratio Z.) At point *q* talc disappears and the actinolite has grown richer in Mg. At point *a* a diopside of composition *b* (and calcite) occurs. I, II, III and IV represent approximately the green schist, the epidote amphibolite, the ampibolite, and the granulite facies respectively (see also p. 66.)

Fig. 9.

$$7\,H_2O + 6\,\underset{\text{tremolite}}{Mg_5Ca_2Si_8O_{22}(OH)_2} \rightleftarrows 5\,\underset{\text{talc}}{Mg_6Si_8O_{20}(OH)_4} + 12\,\underset{\text{calcite}}{CaCO_3} + 8\,SiO_2 \quad \ldots\ldots\ldots\ldots \quad (11)$$

Because of the inability of talc to accept Fe for Mg in its lattice, the $Fe^{\cdot\cdot}$-component of the actinolite series cannot react in the same way.

The assemblage: talc, calcite, and quartz together must be regarded as the low temperature "modification" of tremolite. Talc, calcite, and quartz must accordingly exist in stable equilibrium with gradually $Fe^{\cdot\cdot}$ -richer actinolite as the temperature decreases below the reaction point of equation 11. The equation of this equilibrium is:

$$\text{Mg-rich actinolite} + H_2O \underset{\text{high T}}{\overset{\text{low T}}{\rightleftarrows}} \text{talc} + \text{Fe-rich actinolite} + \text{calcite} + \text{quartz} \quad (12)$$

It is obvious that the equilibrium between a variable actinolite and the other above-mentioned minerals is only dependent on the P,T-conditions if all the minerals: talc, calcite, and quartz are present in the paragenesis.

In rocks carrying all the mentioned minerals the composition of the actinolite may represent a suitable geological thermometer and barometer in the lowest part of the epidote amphibolite facies.[1] Fig. 9 shows the sub-solidus equilibrium diagram of this association.

Example 4. The equilibria of the Al-carrying hornblende and of augite.

When Al replaces Si or (Mg,Fe) in the actinolite-tremolite molecules, the so-called common hornblende appears.

In accordance with the view of different investigators,[2] the lattice of hornblende is able to take up more Al the higher the P,T-conditions. Now we can consider the variation of Al in hornblende from the same view-point as is used in the preceding examples of mixed crystals.

The common hornblende may be looked upon as a solid solution of the pure Al-free actinolite-tremolite molecules and the most Al-saturated “actinolite-tremolite” molecules. For convenience the $Fe^{\cdots}$ and Na-contents are disregarded.

As the Al-rich end component of the series mentioned the common hornblende with the greatest known Al-content must be chosen. The Al-richest varieties of the common metamorphic hornblende lie approximately on the line between (Fe,Mg)-garnet and epidote or anorthite in the ACF-diagram. These Al-end members of the hornblende series will therefore be formed by reaction between garnet and epidote (or anorthite). The Al-richest end components are only produced at the highest P,T-conditions within which hornblende is stable. At lower metamorphic grade the Al-richest hornblende component is unstable and splits up into epidote (anorthite), and almandine-pyrope garnet in accordance with the following equation:

$$\underset{\text{epidote}}{3\,Ca_2\,Al_3\,Si_3\,O_{12}\,OH} + \underset{\text{garnet}}{5\,(Fe,Mg)_3\,Al_2\,Si_3\,O_{12}}$$

$$+ H_2O \rightleftarrows \underset{\text{hornblende}}{3\,Ca_2\,(Fe,\,Mg)_5\,(Al_{6,3}\,Si_{1,7})\,O_{22}\,(OH)_2} \ldots . \qquad (13)$$

It is obvious that the other end components of the hornblende series, the pure actinolite-tremolite molecules, cannot react in the same manner. Hence there must exist an eutectic sub-solidus equilibrium

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[1] H. Ramberg: Loc. cit. p. 96—100.

[2] P. Eskola: Loc. cit. p. 356 and Th. Vogt. Norges geol. unders. Nr. *121* p. 407—410 1927.

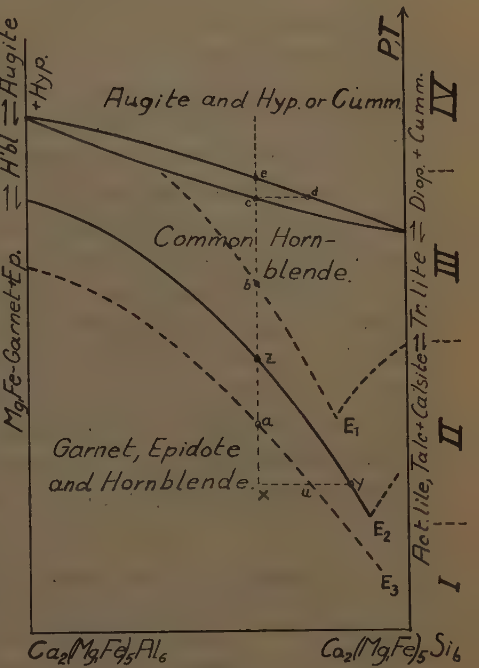

Fig. 10.

Fig. 10. Sub-solidus equilibrium diagram of the common hornblende and augite. It is the ismorphism between Si and Al in the hornblendes and pyroxenes which is of importance for this diagram. The effects of the Mg/Fe-ratios are only slightly indicated by the stipled curves. The solid line from E_3 up to the left is the boundary line between the common hornblende-field and the garnet, epidote-hornblende-field at intermediate Fe/Mg-ratio. The two stipled lines from E_1 and E_3 up to the left are the boundary lines between the mentioned stability fields at Fe/Mg-ratios equal to ∞ and 0, respectively. By great Mg/Fe-ratio there exists an eutectic point at E_1 because of the fact that pure tremolite splits up into calcite and talc, and that talc is not able to accomodate Al in the lattice. The eutectic point gradually moves towards the right and disappears by great Fe/Mg-ratios of the paragenesis. At point x an actinolitic hornblende of composition y exists in association with epidote and garnet by intermediate Fe/Mg-ratios. (By Mg-rich environment only garnet and epidote without hornblende exist at point x, by Fe-rich environment garnet and epidote exist together with common hornblende of composition u). By increasing metamorphic grade, the Al-content of the hornblende which is in equilibrium with epidote and garnet (or chlorite) increases; the composition of the hornblende follows the curve Y Z. At point Z the hornblende is able to accomodate the whole Al-content of the considered assemblage, and epidote and garnet disappear. Between point Z and c the composition of the hornblende is independent of the physical conditions. At point c the stability field of augite is reached and the hornblende splits up in an augite of composition d and hypersthene or cummingtonite. The points a and b are equivalent to Z at pure Fe˙˙ or Mg-parageneses respectively. I, II, III and IV represent the same mineral facies as described in figs. 8 and 9. (See also p. 70—73.)

between the assocation: Fe, Mg-garnet, epidote (anorthite) and a common hornblende with varying Al-content. And, since the minerals: epidote and garnet together represent the low- P,T "modification" of the Al-hornblende end member, the hornblende which is in equilibrium with almandine-pyrope and epidote (An) must gradually get poorer in Al as the P,T decrease below the transformation point of equation 13. These conditions are schematically shown in fig. 10. The Fe/Mg-ratio of the paragenesis has been disregarded in spite of its great importance to the mentioned equilibrium. Mg-rich environments may probably raise the reaction point of equation 13, great Fe-amounts, on the contrary, will lower it (the stipled lines on the fig. 10).

The upper part of the diagram 10 shows the equilibrium conditions of the augite-diopside series. It demands an explanation. In an earlier paper I have been able to prove that the upper P,T-border of the stability field of pure actinolite-tremolite lies lower than the upper P,T-borderline of Al-richer hornblende.[1] Tremolite and actinolite split up into diopside and cummingtonite at the same metamorphic grade at which an Al-rich common hornblende is stable. But if the metamorphism is high enough, even the Al-rich hornblende is unstable and reacts forming augite and hypersthene. The equilibrium between common hornblende and augite is accordingly of the isodimorphic type broadly discussed on p. 50.

Example 5. The biotite—feldspar, garnet equilibrium.

The common paragenesis in granitic gneisses and in granite pegmatites : potash feldspar, garnet and biotite (muscovite) is an illustration of the sub-solidus eutectic transition with reaction products.

Since biotite is stable in this paragenesis it is not easy to see why potash feldspar and garnet do not react forming biotite according to the equation:

$$\underset{\text{garnet}}{(\mathrm{Fe, Mg})_3\,\mathrm{Al_2\,Si_3\,O_{12}}} + 2\,\underset{\text{feldspar}}{\mathrm{KAl\,Si_3\,O_8}} + 2\mathrm{H_2O} \rightleftarrows$$

$$\underset{\text{biotite}}{\mathrm{K\,(Fe, Mg)_3\,Al\,Si_3O_{10}\,(OH)_2}} + \underset{\text{muscovite}}{\mathrm{KAl_3\,Si_3\,O_{10}\,(OH)_2}} + 3\,\mathrm{Si\,O_2} \; . \; . \; . \; . \qquad (14)$$

But since the garnet in pegmatites and granites according to general experience[2] are spessartine — almandine varieties, the paragenesis: garnet, potash feldspar, biotite does not indicate any incomplete equilibrium.

In common biotites from granites and pegmatites, Mn cannot replace Fe·· or Mg to any considerable extent. Because of this fact, pure Mn-garnet cannot react with potash feldspar forming biotite.

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[1] H. Ramberg loc. cit. p. 105.

[2] W. I. Wright: Am. min. *23.* p. 436. 1938.

The garnets of granites and pegmatites are thus approximately two-component mixed crystals: spessartine—almandine, of which only the almandine component can react with feldspar, giving biotite.

It seems to be empirically proved that the biotite is stable at lower temperature and pressure than the combination: almandine—potash feldspar, i. e. that the right side of equation 14 is stable at low, and the left side at high temperature and pressure. Accordingly there must exist a varying equilibrium between biotite, potash feldspar, and $Fe^{\cdot\cdot}$, Mn garnet. In this equilibrium the garnet must gradually grow richer in spessartine as the temperature decreases below the reaction point of equation 14.

Only when potash feldspar occurs in the presence of quartz and water the equilibrium between biotite and garnet is determined by the P,T-conditions regardless of the chemical environment. In the absence of potash feldspar (which is the visible proof of potassium-excess) even the Mn-free garnet may occur together with biotite at low temperatures, the amount of potassium in the paragenesis being insufficent to form biotite from the almandine. Disregarding Mg the equilibrium equation may be written:

$$\overset{\text{high T.}}{\text{Mn-poor (Fe, Mn)-garnet} + \text{K-feldspar} + H_2O} \rightleftarrows \overset{\text{low T.}}{\text{Mn-rich garnet}} + \text{biotite} + \text{muscovite} + \text{quartz} \;\ldots\ldots\ldots\ldots \quad (15)$$

(Commonly biotite is so rich in Al that muscovite does not appear in equations 14 and 15.)

In the granulite facies even pure Fe, Mg-garnet may exist in contact with potash feldspar without forming biotite. At that high temperature and pressure biotite is probably unstable. The transition point of equation 14 consequently coincides with the lowest borderline of the P,T-field represented by the granulite facies. The Mn/(Fe, Mg) ratio of the garnet which occurs in equilibrium with biotite and potash feldspar thus represents a geological P,T-indicator valid in the interval between granulite facies and the stability field of low grade pegmatites and gneisses, (epidote amphibolite facies). According to the literature[1] the petrographic experience supports this theoretical explanation. The garnets which are found in the low temperature clevlandite-quartz pegmatites are often approximately pure spessartines. The granulite garnets, on the contrary, are usually very low in Mn. In

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[1] W. I. Wright: Am. min. *23* p. 436. 1938.

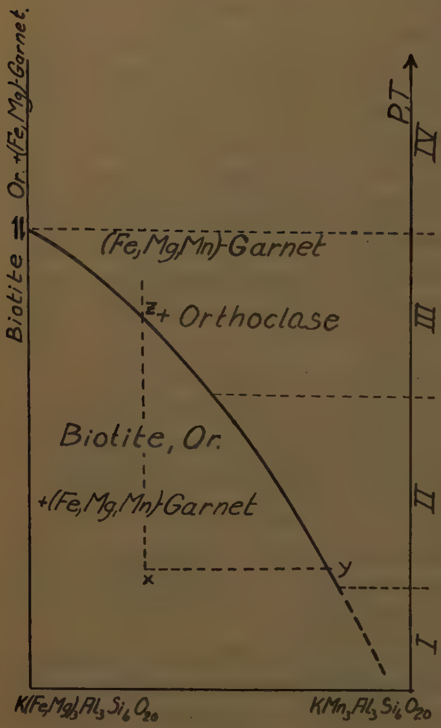

Fig. 11.

Fig. 11. The equilibrium diagram of biotite, potash feldspar, and spessartine-almandine garnet. The relation between the different elements in the parageneses which the diagram represents is seen from the fig. SiO_2, (as quartz), Al_2O_3 (as muscovite) can be present in excess. H_2O must be present, and CaO may also occur in plagioclase without displacing the equilibrium. It is the Mn/(Mg, Fe) ratio which is of importance for the diagram. At point *X* exists a garnet of the Mn/(Fe, Mg) ratio *Y* in equilibrium with potash feldspar and biotite. By increasing metamorphism the Mn/(Fe, Mg) ratio of the garnet follows the curve *Y Z*: the garnet grows richer in Fe and / or Mg, concomitantly the amount of biotite decreases. At point *Z* all the biotite is split up to garnet and feldspar. I, II, III and IV represent the same facies as in fig. 7—8. (See also p. 70—73.)

gneisses carrying potash feldspar, and in granites, the spessartine component in the garnet varies from low to high, these rocks being formed at different P,T-conditions.

Fig. 11 is an equilibrium diagram showing the stability conditions of the garnet paragenesis. For convenience the effect of the Fe/Mg ratio on the equilibrium is disregarded.

Example 6. The orthohypersthene—clinohypersthene equilibrium.

The series: orthorhombic hypersthene — clinohypersthene is a typical isodimorphic series. The monoclinic series consists of the high temperature modifications.

For the pure Mg-component, the enstatite, the transition temperature is about 1140°; and for the Fe-richest hypersthene investigated, the transition temperature is about 995° according to Bowen and Schairer.[1]

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[1] Am. Jour. Science, XXIX, p. 151—217. 1935.

It is of great interest that Bowen and Schairer have been able to show experimentally the transformation in sub-solidus phase between monoclinic and orthorhombic hypersthene. According to their experiments and equilibrium diagram[1] a very small two phase interval seems to exists between the monoclinic and rhombic members. But there is some indication that the perfect sub-solidus equilibrium of the pyroxenes of intermediate composition was not developed during Bowen and Schairer's experiment. For the pure Mg-enstatite and the Fe-richest hypersthene the transition is very sharp at 1445° and 995° respectively. For the intermediate bronzite members, however, the transition is unprecise; the orthorhombic bronzites could be superheated by 300° above their transition temperature. But by using NaF as a catalyst the transition of the brozites became distinct. "In working with added NaF it was found that some monoclinic crystals would form from orthorhombic even at low temperature where we had every reason to believe that the orthorhombic was the stable form." Above 1120° an orthorhombic bronzite of definite composition recrystallized to clinobronzite only. "But below 1120° newly formed crystals of bronzite appear together with the clino-bronzite, which proves the stability of the orthorhombic bronzite at temperatures below 1120° and places the transformation temperature at 1120°."

Bowen and Schairer assumed that the clinobronzite which appears together with the orthorhombic brozite was unstable; it was formed below its stability temperature. Because of this assumption Bowen and Schairer thought the immiscibility field between orthorhombic and clino-hypersthenes to be very small.

It seems natural to me, however, that the fact that both orthorhombic and monoclinic bronzites were formed together at temperatures below 1120° indicates that there exists a larger transformational field between the two pyroxenes than assumed by Bowen and Schairer.

But it is reasonable to think that the perfect chemical equilibrium was not developed at the transformation of the bronzite members. In the pure end component the transition between the rhombic and monoclinic modification may be explained as a distortion of the lattice analogous to the transition: α-quartz $\rightleftarrows$ β-quartz. No chemical diffusion between the two different phases is necessary. In the bronzitic crystals, however, the compositional difference of the two phases which are in stable equilibrium demands a chemical diffusion between the two phases. Mg and Fe must migrate from one crystal to another. Without NaF as a catalyst this chemical reaction was not able to take place in the solid phase. But it is of course not evident that the added NaF in the experiment of Bowen and Schairer was able to develop the complete equilibrium in the solid phase.

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[1] Bowen and Schairer, loc. cit.

Similar isodimorphic transitions are found by other equilibrium experiments also. Greig and Barth[1] have for instance proved the existence of a two phase field between carnegieite and nephelite.

In nature clinohypersthene (pigeonite) only occurs in basic lavas and meteorites. According to the rapid cooling of lavas a perfect development of the equilibrium is not probable. The miscibility between clinohypersthene and diopside in pigeonites surely also influences the equilibrium between clinohypersthene and hypersthene. (p. 68.) The mentioned sub-solidus equilibrium can therefore probably not be used as any suitable P,T-indicator.

In regional metamorphic rocks, in migmatites and perhaps in some plutonic igneous rocks the equilibria between the minerals are well on the way to complete attainment. Isodimorphic series among metamorphic minerals may therefore give usable geological thermometers and barometers. Ideal isodimorphic series, however, are not common among metamorphic minerals. But in the case where the fields of stability of two isomorphic series: (A, B) X and (A, B) Y where X and Y differ more or less, border against each other, the equilibria conditions are analogous to the conditions in isodimorphic series, (see p. 52 equations 7 and 8).

Example 7. The orthohypersthene—anthophyllite equilibrium.

Bronzite: $(Mg, Fe)_2 Si_2 O_6$, and anthophyllite: $(Mg, Fe)_7 Si_8 O_{22} (OH)_2$ are two isomorphic series of the above mentioned type. The difference in composition of the two minerals is made up by the H_2O-content and the greater Si/(Mg, Fe) ratio of the hornblende. But since both minerals are saturated in silica, and the pyroxene may probably be stable in the presence of H_2O, we may consider the system with excess of quartz and water. Then the transformation: pyroxene $\rightleftarrows$ anthophyllite is in principle analogous to the transformation between ideal isodimorphic series.

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[1] J. W. Greig, T. F. W. Barth: Am. J. Sci. XXXV A, p. 94. 1938.

There is every reason to believe that the transformation temperature of the pure Fe-members is lower than that of the pure Mg-members, in analogy with most other Fe-Mg silicates. Accordingly the sub-solidus equilibrium diagram is schematically shown in fig. 8 (p. 54).

Orthorhombic pyroxene is a critical mineral for the granulite facies in regional metamorphic rocks. Antophyllite is supposed to be unstable at P,T-conditions higher than those of the amphibolite facies. From fig. 8 we see that the transformation temperature: hypersthene $\rightleftarrows$ antophyllite varies with varying Fe/Mg ratio. Fig. 8 further indicates that both pyroxene and hornblende may exist side by side in stable equilibrium (in excess of quartz and water), by definite correlations between P,T and Fe/Mg ratios. In accordance with this consideration the P,T-borderline between granulite and amphibolite facies will be very indistinct if based on the transformation: orthorhombic pyroxene $\rightleftarrows$ anthophyllite.

The equilibrium conditions become complicated by the occurrence of monoclinic cummingtonite at temperatures below the stability field of bronzite.[1]

Bowen and Schairer's synthesis of fluorine cummingtonites is in good accordance with the explanations above.[2]

Orthorhombic pyroxenes of different compositions were mixed with NaF and heated. The products were fluorine cummingtonites of different compositions. In harmony with the theoretical, diagram fig. 8, the newly formed hornblendes always had greater Mg/Fe ratios than the pyroxenes (tab. I). In spite of the possibility that the complete equilibrium was not attained, such differences in the Mg/Fe ratio of pyroxenes and hornblendes at least indicate a higher transformation temperature for Mg-members than for Fe-members, and the existence of a transformation field between the two phases.

Table I

Pyroxene Mol.% Mg-silicate	Hornblende Mol% Mg-silicate
100	100
84	87
68	73
47	62
26	48
0	0

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[1] A. N. Winchell: Am. min. *20*, p. 329; 1935.

[2] Am. min. *20*, p. 543, 1935.

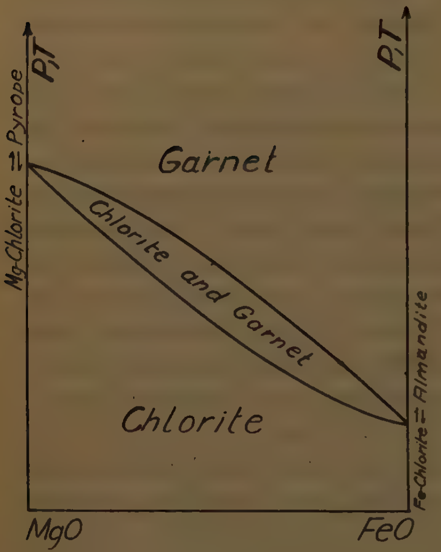

Fig. 12.

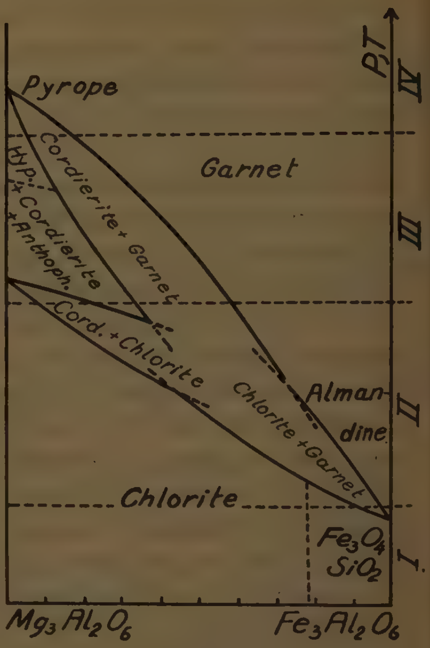

Fig. 13.

Fig. 12. Equilibrium diagram of the chlorite-garnet association.

Fig. 13. Equilibrium diagram of the (Mg, Fe)-alumosilicates in sub-solidus state. The ratio (Fe, Mg)/Al is $^3/_2$. All the minerals are saturated in silica so that quartz can occur all over the diagram without any distortion effect. H_2O must be present. I, II, III and IV represent the same facies as in fig. 7—8, (see also p. 70—73).

Example 8. The chlorite—garnet equilibrium.

In low grade schists the paragenesis: chlorite—almandine is often encountered. The ratio (Mg, Fe)/Al is approximately equal in both minerals. But the relative contents of Si of the two mineral series differs, and in contradistinction to garnet chlorite contains H_2O.

In the presence of quartz and water the association chlorite and almandine must either be explained by chemical unstability or by the effect of the Mg/Fe ratios of the parageneses.

It is of great significance that the garnets in low grade rocks always are Mg-poor almandine while the chlorites in the same rocks are relatively richer in Mg. Exactly this difference in the Mg/Fe ratios of chlorite and almandine will appear if Mg-chlorite transforms to pure pyrope at higher temperatures than Fe-chlorite transforms to pure almandine. Fig. 12 shows the conditions at the phase transition: chlorite $\rightleftarrows$ (Fe,Mg)-garnet. The equilibrium diagram is probably not valid at the Mg-end. Pyrope is stable only at high P,T-conditions, and it does not seem probable that chlorite is the mineral which occurs immediately beneath pyrope in the P,T scheme (page 64 fig. 13).

Before leaving the garnet—chlorite association, we shall look at the extreme low temperature parageneses containing spessartine-rich garnets.

In some regional metamorphic provinces garnet appears before the biotite isograde. Empirically these low temperature garnets are richer in spessartine than the usual almandines occurring in and above the biotite isograde.[1]

Physico-chemically the paragenesis may be explained by the fact that the presence of Mn suppresses the partial gas tension of Fe and Mg above the garnet. The (Fe, Mg) gas tension of the garnet may thus be lower than the (Mg, Fe) gas tension of chlorite even at low temperature. In this way manganese will stabilize garnet at temperatures lower than usual. In other words: there exists a eutectic equilibrium between chlorite and (Mn, Fe, Mg)-garnet. The chlorite is the low temperature modification of the (Fe, Mg)-garnet; but the Mn-garnet has no low temperature modification which is miscible with chlorite.

Example 9. The garnet—cordierite equilibrium.

The association: cordierite— (Fe, Mg)-garnet occurs in some gneisses and granulites. The chemical compositions of cordierite and garnet allow the two minerals to exist in stable equilibrium at certain (Fe, Mg)/Al-ratios in the paragenesis regardless of the Fe/Mg-ratio. But cordierite and garnet also occur together in rocks in which the (Fe, Mg)/Al-ratio is different from the above mentioned permissible values. Such parageneses are to be explained by the Fe/Mg-ratios.

In Fe-rich rocks almandine occurs, in Mg-rich rocks cordierite.[2]

The association garnet, cordierite and anthophyllite is common. Anthophyllite is the pure (Fe, Mg)-silicate which is stable at the same P,T-conditions as cordierite and garnet. The pyrope-component of the garnet may be assumed to form from pure Mg-cordierite and Mg-anthophyllite as shown by the equation:

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[1] A. Harker: Metamorphism. London 1932.

[2] P. Eskola, in Barth, Correns, Eskola, loc. cit. p. 354.

$$\underset{\text{cordierite}}{21\,Mg_2Al_4Si_5O_{18}} + \underset{\text{anthophyllite}}{12\,Mg_7Si_8O_{22}(OH)_2} \rightleftarrows \underset{\text{pyrope}}{42\,Mg_3Al_2Si_3O_{12}} + 12\,H_2O \quad (16)$$

Pure Fe-cordierite is probably unstable, but cordierite containing about 60 mol. per cent Fe-silicate is known.[1] Pure almandine may theoretically be formed from Fe-anthophyllite and a hypothetical Fe-cordierite:

$$\underset{\text{Fe-cordierite}}{21\,Fe_2Al_4Si_5O_{18}} + \underset{\text{Fe-anthophyllite}}{12\,Fe_7Si_8O_{22}(OH)_2} \rightleftarrows \underset{\text{almandine}}{42\,Fe_3Al_2Si_3O_{12}} + 12\,H_2O \quad (17)$$

The garnet-side of the equations is probably realized at high pressure; the cordierite side, on the contrary, is the high temperature side. (Under ordinary pressure cordierite melts directly without forming pyrope.)

Under the P,T-conditions of regional metamorphism it seems as if the effect of pressure exceeds the effect of temperature. Accordingly the right hand sides of equations 16 and 17 are stable at high grade and the left hand sides at lower grade of metamorphism. Further we suppose a higher transformation point of equation 16 than of 17.

In nature these schematic conditions are complicated by the above discussed equilibria between garnet and chlorite, and between chlorite and cordierite. Taking these facts into account we obtain a diagram as shown in fig. 13. Fig. 13 explains the equilibrium condition between (Fe, Mg, Al)-silicates during regional metamorphism approximately.

Example 10. The actinolite—diopside equilibrium.

In the two series: actinolite and diopside, similar equilibrium conditions must exist at definite (Fe, Mg)/Ca-ratios. There is a considerable difference in the (Fe, Mg)/Ca-ratios in the actinolite and the diopside series. In quartzous rocks saturated with H_2O the two minerals may therefore occur together in stable equilibrium within a relatively great (Fe, Mg)/Ca-interval, regardless of the Fe/Mg-ratio.[2] But actinolite and diopside are also found together at an average chemical composition different from the permissible ones. In some marbles, for instance, actinolite and diopside occur in contact with calcite. But in that case the actinolite is often richer in Fe than the diopside.[1]

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[1] A. Brammall and B. Rama Rao, Min. Mag. XXIV p. 257, 1936.

[2] P. Eskola: The Mineral facies of rocks. Norsk geologisk tidsskrift *6*, 1921

For the transformation: actinolite ⇄ diopside we may put forth the equations:

$$\overset{\text{low T.}}{SiO_2 + \underset{\text{actinolite}}{CaFe_5Si_8O_{22}(OH)_2} + 3CaCO_3} \rightleftarrows \overset{\text{high T.}}{\underset{\text{hedenbergite}}{5CaFeSi_2O_6} + H_2O + CO_2} \quad (18)$$

$$SiO_2 + \underset{\text{tremolite}}{CaMg_5Si_8O_{22}(OH)_2} + 3CaCO_3 \rightleftarrows \underset{\text{diopside}}{5CaMgSi_2O_6} + H_2O + CO_2 \quad (19)$$

where the transformation point for eq. 19 lies higher than for eq. 18. The upper part of diagram 9 results from the explained condition, and from the petrographic experience that the pyroxene side of the equations is stable at higher metamorphic grade than the actinolite side.

Example 11. The phlogopite—calcite, microcline equilibrium.

The characteristic mica of the metamorphic limestones is phlogopite. The mineral occurs in both contact and regional metamorphic marbles. Because of the content of fluorine in phlogophite, it is very reasonable to assume a pneumatolytic origin of the phlogopites in metamorphic limestones. But the occurence of Mg-mica in ordinary regional metamorphic skarns without other pneumatolytic minerals demands a different explanation of its origin.

T. F. W. Barth has suggested that the occurrence of metamorphic phlogopite in marbles is due to a deficit of silica in the paragenesis.[2] But this assumption cannot always explain the fact that the Fe-rich biotite does not appear in intermediate and high grade limestones. Biotite only occurs in low grade marbles where hornblende and pyroxene are not stable and hence biotite will not react whith calcite forming hornblende and diopside. But as soon as the P,T-conditions cross the lowest stability boundary of hornblende or diopside, biotite disappears leaving its place to lime silicates in contact with calcite.

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[1] H. Ramberg loc. cit. p. 61.

[2] T. F. W. Barth: N. Jahrb. f. Min., B. Bd. 57, Abt. A, p. 1103, 1928.

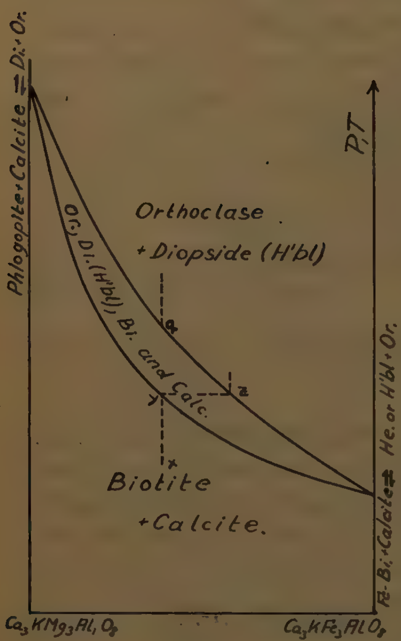
Phlogopite+Calcite ⇄ Di.+Or.
P,T
Orthoclase
+Diopside (H'bl)
Or.+Di.(H'bl), Bi. and Calc.
a
z
y
x
Biotite
+Calcite.
Fe-Bi.+Calcite ⇄ He. or H'bl+Or.
$Ca_3KMg_3Al_1O_8$
$Ca_3KFe_3AlO_8$

Fig. 14.

Fig. 14. Sub-solidus equilibrium diagram of the biotite, calcite ⇄ potash feldspar, lime silicate parageneses. Other than the elements which are given in the fig., silica and water must be present in sufficient amount.

The Mg-rich phlogophite, on the other hand, may occur up to the highest grade of metamorphic marbles.

The biotite series will react with calcite forming lime silicates (hornblende or diopsidic pyroxen) and potash feldspar. The feldspar, hornblende side of the equation is the high temperature paragenesis. But we must assume that the pure Fe-end member reacts at considerably lower temperatures than the pure Mg-end member of the phlogopite-lepidomelane series. According to this explanation the diagram fig. 14 appears. It gives a simple and natural explanation of the occurrence of phlogopite and the absence of biotite in metamorphic marbles.

Example 12. The hypersthene—pigeonite equilibrium.

As the last example of sub-solidus transition we shall consider a combination of the eutectic and the isodimorphic type.

The pyroxene of the basaltic lavas, the pigeonite, is a typical illustration of such a combination type. Disregarding Al, $Fe^{\cdot\cdot\cdot}$ and Na, the pigeonite is a mixed crystal of the clinohypersthene and the diopside series:

$$(Mg, Fe)_2\, Si_2\, O_6 \text{ and } (Mg, Fe)\, Ca\, Si_2\, O_6.$$

The Ca-free compounds occur as isodimorphic series (p. 60) the high temperature modifications of which are monoclinic and the low temperature modifications orthorhombic. Only the high temperature modifications are miscible with the diopside, forming pigeonite. Diopside has only one modification. Accordingly, there must exist a eutectic sub-solidus equilibrium between the orthorhombic hypersthene ($(Mg, Fe)_2Si_2O_6$) and the 4-component mixed crystal: pigeonite

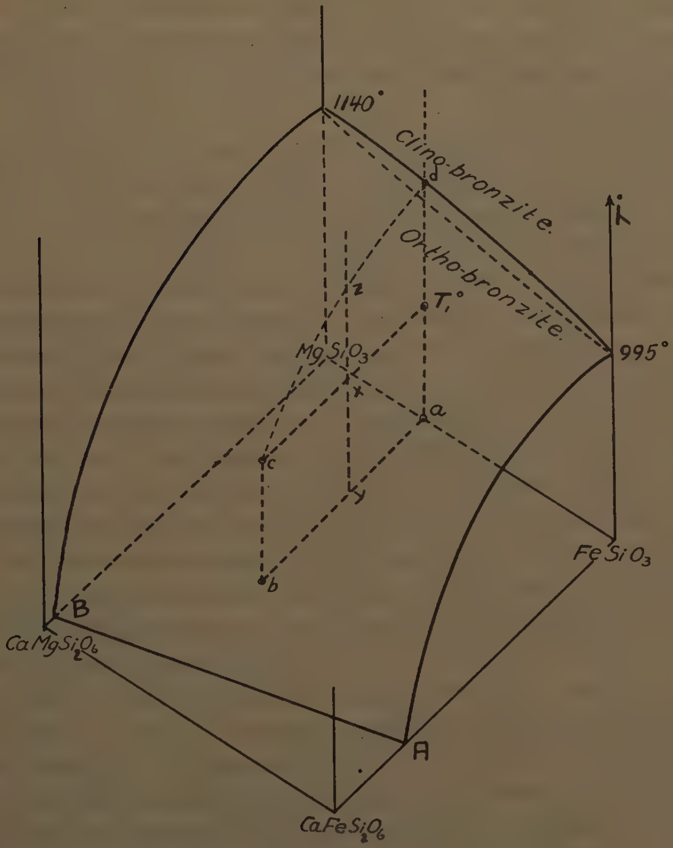

Fig. 15.

Fig. 15. Equilibrium diagram of the orthopyroxene (enstatite) and the clinopyroxene (pigeonite) in solid state. The vertical plane $FeSiO_3$, $MgSiO_3$, 1140°, 995° is redrawn after Bowen and Schairer (op. cit.) The vertical plane rising from $FeSiO_3$, — $CaFeSi_2O_6$ shows the equilibrium diagram of the eutectic sub-solidus phase transition between pure ferrosilite and the mixed crystal clinoferrosilite-hedenbergite. The plane rising from $MgSiO_3$, $CaMgSi_2O_6$ represents the equilibrium of eutectic type between pure enstatite and the mixed crystal clinoenstatite — diopside. Above the curved surface A, B, 1140°, 995° only pigeonite (and diopside) is stable. Below the mentioned surface pigeonite is in stable equilibrium with orthopyroxene. At point X (temperature equal to T_1° and chemical composition given by point Y) a pigeonite of composition *c* (or *b*) exists in equilibrium with rhombic hypersthene of composition *a*. As the temperature increases from X to Z, the pigeonite gradually grows richer in (Fe, Mg)-silicate, and at point Z the last rest of the orthopyroxene disappears.

At temperatures decreasing below the transformation point: clinohypersthene $\rightleftarrows$ ortohypersthene the pigeonite must gradually get poorer in the (Mg, Fe) SiO_3 compound, which separates from the pigeonite as orthohypersthene.

But the Ca/(Mg, Fe) ratio of the pigeonite which is in stable equilibrium with orthohypersthene is not only dependent on the P,T-conditions; also the Fe/Mg-ratio in the pyroxenes play a role. Fig. 15 shows schematically the equilibrium conditions.

Even at temperatures far below the transformation point of the Ca-free pyroxenes, the pigeonitic 4-component mixture must exist, but now with a smaller amount of the $(Mg, Fe)_2Si_2O_6$-compound in solid solution. We have a diopsidic pyroxene with excess of Mg and Fe.

The association: orthopyroxene — clinopyroxene (pigeonite) has been discussed by many investigators (see Barth[1] and the references in his paper) but always with the conclusion that the pigeonitic pyroxenes which occur in rocks containing orthopyroxenes are metastable phases. In my opinion the two pyroxenes are, as we have seen, stable together over a wide temperature range.

Diopside and hypersthene is a common association in rocks of the charnocite-anorthosite kindred, and in rocks belonging to the granulite facies. An excess of the hypersthene-compound in the diopside of these rocks is very common.[2]

Under the high P,T-conditions represented by the granulite facies the equilibrium between the minerals probably is well on the way to complete adjustment. The excess of Mg- and Fe-silicate in the diopside existing together with hypersthene may accordingly appear as a usable geological thermometer and barometer within the granulite facies.

IV. THE SUB-SOLIDUS REACTIONS IN MIXED CRYSTALS, AND THE FACIES CLASSIFICATION OF ROCKS

A definite mineral facies represents a definite P,T-interval within which the rocks have crystallized or recrystallized. The different facies are characterized by the stability field of their critical or typomorphic minerals. Since these index minerals usually are solid solutions of common rock-forming minerals the phase-transitions of solid solutions, as discussed above, are of great importance in establishing the border lines between the several facies.

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[1] T. F. W. Barth: Am. J. Sci. 1936, p. 336—339.

[2] A. F. Buddington: Geol. Soc. Memoir 7, p. 39. 1939.

We shall now briefly go through the normal regional metamorphic mineral facies.[1]

1. The transition: green schist facies $\rightleftarrows$ epidote amphibolite facies.

In the green schist facies the association: *talc—calcite* is stable. By crossing the P,T-border line against the epidote amphibolite facies, talc and calcite react forming actinolite (p. 55). But this transformation "point" varies with the Fe/Mg-ratio in the paragenesis. In Fe-rich environments the transformation point lies lower than in Mg-rich environments; i. e. at high Mg/Fe-ratio the green schist paragenesis: talc-calcite exists at higher temperatures than does the amphibolite facies index mineral: actinolite at lower Mg/Fe-ratio, (fig. 9). The whole isomorphous actinolitic series is therefore not suitable as a basis for distinguishing between the two facies. The Fe/Mg-ratio of the actinolite which shall characterize the transition: green schist facies $\rightleftarrows$ epidote amphibolite facies must be defined.

The transformation: *talc* $\rightleftarrows$ *anthophyllite* lies between the two above mentioned facies. This transition is also highly dependent on the Fe/Mg-ratio, (fig. 8 p. 54).

The transformation: *chlorite* $\rightleftarrows$ *almandine* probably lies at the P,T-condition under consideration. Here the Fe/Mg-ratio is not the only factor; also the Mn-contents of the paragenesis are of the greatest importance, as shown on p. 64.

2. The transition: epidote amphibolite facies $\rightleftarrows$ amphibolite facies.

According to Eskola,[2] the epidote amphibolite facies is charaterized by unstability of any amount of anorthite in the plagioclase. All the "anorthite molecule" exists in the epidote minerals. These conditions, however, occur only at very low temperature; surely far below the temperature which Eskola thougt to be the highest stability temperature of the epidote amphibolites — the rocks which have given the facies its name.

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[1] P. Eskola, in Barth, Correns and Eskola loc. cit. p. 334—360.

[2] P. Eskola, Loc. cit. p. 357.

I will therefore propose to fix the lowest P, T-boundary of the epidote amphibolite facies by the association: *Epidote* — $An_{10}\,Ab_{90}$, and the highest P, T-boundary by the association: *Epidote*— $An_{80}\,Ab_{70}$. Below An_{10}, I propose to put green schist facies, and above An_{80} we put amphibolite facies which reaches up to the highest P, T-values at which common hornblende is stable.

3. The transition: amphibolite facies ⇄ granulite facies.

The association *pyrope-almandine-garnet* and *potash feldspar* is critical for the granulite facies. On crossing the borderline into the amphibolite facies the two mentioned minerals react, forming biotite (p. 58—60). According to the explanations on p. 59 the reaction point is highly influenced by the Mn/(Fe, Mg) ratio of the paragenesis. Some amount of Mn in the garnet stabilizes the association: garnet—potash feldspar at P,T conditions lower than usual.

Another critical transformation between the two last mentioned facies is the reaction : *bronzite* ⇄ *anthophyllite* (or cummingtonite). This transformation is also dependent on the Fe/Mg-ratio of the rocks (p. 62).

4. The transition: sanidinite facies ⇄ hornfels facies.

Pigeonite is a critical mineral in the sanidinite facies (diabase facies). By the transition to hornfels facies, or to granulite facies also, the pigeonite decomposes, forming diopsidic pyroxene and hypersthene. As discussed in example 12 p. 64 the decomposition point varies greatly with the Ca/(Fe, Mg)-ratio and also with the Fe/Mg-ratio of the paragenesis.

The importance of the eutectic and the isodimorphic types of sub-solidus phase transition for the facies classification of rocks is best understood by considering the equilibrium diagrams figs. 6—15.

Of course, one must remember that most of the diagrams are more or less hypothetical. They give the principal features of the sub-solidus equilibrium, the exact mineralogical and chemical data can only be found by very accurate paragenetical and chemical studies of metamorphic rocks.

Another thing of the greatest significance in connection with the diagrams shown in this paper is the mutual dependence of pressure and temperature during the metamorphism. It seems as if there exists a normal regional metamorphism during which P and T vary sympathically. High grade metamorphism therefore means that both P and T are high, low grade that both P and T are low.

The equilibrium diagrams represent conditions during this normal regional metamorphism. The fields I, II, III and IV in the different diagrams represent the green schist facies, the epidote amphibolite facies, the amphibolite facies and the granulite facies respectively. And since every mineral facies has its definite P,T-field it is obvious that the borderlines between the different facies cannot coincide with reaction or transformation points in all diagrams. In diagram 9, for instance, the transition: talc $\rightleftarrows$ Mg-tremolite coincides with the boundary between epidote amphibolite facies and amphibolite facies. In diagram 7 the same boundary represents no discontinuous point, but is characterized by the association epidote—$An_{85}Ab_{65}$, and so on.

Acknowledgements.

The writer thanks Professor dr. Tom. F. W. Barth for critically reading the manuscript and correcting the English. Thanks are also due to konservator dr. Ivar Oftedal for friendly criticism of the manuscript.

Mineralogisk Institutt Oslo, Sept. 1944.

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HIGH TEMPERATURE OPTICS IN PLAGIOCLASES OF THE OSLO REGION

BY

CHRISTOFFER OFTEDAHL

With 1 figure in the text.

The exact knowledge of the variation of the optical orientation with the chemical composition in the plagioclase series, is principally due to the work of Becke. He used the conoscope for his determinations, but later the universal stage gradually came into use, and the methods for this are given by L. Duparc and M. Reinhard (1924, in French) and by M. Reinhard (1931, in German). The stereograms with the migration curves were suited for most of the plagioclases, but in some cases the literature shows that there were obvious discrepancies. The probable reasons for this may be ascribed to various causes: The potash content, wrong migration curves, unexact orientation of the twins, errors of measurement, etc. A. Köhler (1942a) is the first who realized that most of the abnormal feldspars were derived from volcanic rocks; indeed all plagioclases from such rocks deviate from the migration curves of the previous authors. Köhler speaks of the existence of a "Hoch- und Tieftemperaturoptik". To show the properties of this new high temperature series, Köhler has worked out a new diagram type for the migration curves. The values of the angles $\alpha_1\alpha_2, \gamma_1\gamma_2$ and of the angles between the different optical axes for the common twins are given as a function of the An content. The curves of these "Köhler values" are given for the whole low temperature series and the interval 30—60 An of the high temperature series.

The true existence of such a series is proved by H. Scholler (1942), who transformed low temperature plagioclases into the high temperature type by exposing them to heat. The results of former investigations of similar kind are now easy to understand. T. F. W. Barth (1931) succeeded in transforming low temperature plagioclases into the high temperature type, but C. T. Barber (1936) who treated high temperature feldspars, did not find any noticeable change.

Table 1.

No.	Rock	Angles		High temp. values	Low temp. values	Mean of high temp. values	Powder determ.	Zone method
1	Rhomb porph. RP₁₃ d Sørkedal	$\alpha\ \alpha^{\pi}$ $\gamma\ \gamma^{\pi}$ $A\ A^{\pi}$	7 115 70	43 40 44	40 36 47	42	42	54
2	— » —	$\alpha\ \alpha^{\pi}$ $\gamma\ \gamma^{\pi}$ $A\ A^{\pi}$	10 117 68	45 51 45	42 51 49	44		50
3	Rhomb porph. RP₁₃ b Sørkedal	$\alpha_1\ \alpha'_2$ $\gamma_1\ \gamma'_2$ $B_1\ B'_2$	94 28 54	48 42 46	45 58 48	45	42	50
4	Akerite porph. N of Burudvatn, Bærum.	$\alpha_1\ \alpha_2$ $\gamma_1\ \gamma_2$ $A_1\ A_2$	78 124 67	45 51 51	43 61 56	49	—	63

Four stereograms with their "angle values" of Köhler and the corresponding An values in the high and low temperature series. Nos. 1 and 2 are measurements on different parts of a large crystal.

H. Tertsch (1942) has made measurements on basic synthetic plagioclases and has thereby enlarged the interval of known high temperature values to 30—100 An. A paper on the missing "acid" interval was announced.

For a paper on the petrology of some volcanic rocks in the Oslo region, I wanted to determine the plagioclase phenocrysts in akerite porphyries and rhomb porphyries. Stereograms of akerite feldspars were prepared by me. Most of them did not correspond very well to the migration curves of Duparc and Reinhard, although a few did. Some of them showed a good accordance for the one twin, but not for the other. It was believed that inaccurate measurements were the chief source of the differences, and to save time the zone method of Ebert (1931) was used for the succeeding determinations.

Then some rhomb porphyry phenocrysts, which were large enough and sufficiently unaltered, were determined by immersion liquids, and the determinations showed 7—9 An less than the zone method determinations! By seeking the reason for this, Köhler's paper was

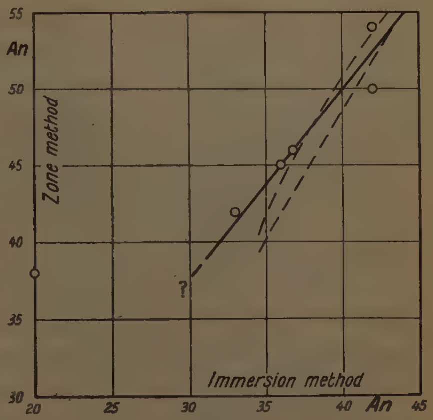

Fig. 1.

The An values of some rhomb porpyhry phenocrysts, found by the immersion method, with the corresponding values from the zone method. The curve should thus give the correspondence between the new correct An values, and the values which are derived in the usual manner: Extinction angles ⊥ (001) and (010) or ⊥ (010). The two broken curves give the same correspondence; they are derived from curves of the extinction angles of Köhler and of Tertsch.

discovered. A reexamination of the stereograms was made, and some new ones were prepared. The angles then available were compared with the curves of Tertsch (1942, p. 206—211) for the two series. Since every angle correponds to two An contents (one for each of the two series), each diagram gives two sets of values, one set corresponding to the high temperature optic, the other set to the low temperature optic. The individual readings of the high temperature set give the most consistent values, but it must be admitted that the consistency is in no way as good as in the examples stated by Köhler (1942 b). Table 1 gives the angles from four diagrams, and their interpretation. The variations within the individual high temperature values do not surpass 6 An, while the low temperature values exhibit much greater variations. Feldspar No. 4 of Table 1 demonstrates that the basic plagioclases get a considerable reduction in their An values by using the high temperature curves; thus 63 An in the phenocrysts of an akerite porphyry (determined on several crystals) is a sensational value, but reduced to 49 An, it becomes a common one.

There are reasons to believe, therefore, that the feldspars listed in Table 1 exhibit the "Hochtemperaturoptik" and not the "usual" plagioclase optics.

This may be checked on specimens which can be determined with the immersion method. This seems not to have been tried by the Austrian authors. Some rhomb porphyry feldspars were sufficiently unaltered to be determined by this method, and the results are plotted against the zone method values. See Fig. 1.

For comparison corresponding curves may be taken out of the diagrams of Köhler and Tertsch for the extinction angles ⊥ (001) and

(010) in the two series. The maximum extinction angles in the zone [010] coincide with the extinction angles ⊥ (001) and (010) up to 55 An, and the last mentioned extinction angles may therefore be compared with values derived from the zone method. The difference in the trend of the broken curves in the figure is caused by the fact that there is some difference in the curves for the extinction angles ⊥ (001) and (010) for low temperature plagioclase in the tables of Köhler, compared with those of Tertsch.

It is to be expected that the phenocrysts of the rhomb porphyry lavas belong to the high temperature series; more interesting is the fact that the insets of an intrusive rock with lava texture (= akerite porphyry) also do so. It is possible, however, that they belong to a modified high temperature series; in this case there is a transitional change in the optics of the high and low temperature modifications; the two feldspars thus having the relation of polytropy, (Barth 1934, p. 283).

Acknowledgement.

The investigation was carried out at the Mineralogical Institute of the University, and I am greatly indebted to the director, professor Tom. F. W. Barth, for kind interest during the work.

Sept. 26th, 1944. Mineralogisk Institutt, Oslo.

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Printed February 26th, 1945.

Ms. rec. Nov. 21, 1944.

ON THE CRYSTAL STRUCTURE OF PHOSGENITE $Pb_2Cl_2CO_3$ AND THE SYNTHETIC COMPOUNDS $Pb_2Cl_2CO_3$, $Pb_2Br_2CO_3$, $Pb_2(Br,Cl)_2CO_3$.

BY

IVAR OFTEDAL

With 5 figures in the text.

About 1930 I spent a considerable amount of time in trying to determine the crystal structure of phosgenite by means of Laue, powder, and oscillation photographs, the material being fragments of a large crystal from Monte Poni, Sardinia, belonging to the collections of the Mineralogisk-geologisk Museum in Oslo. The research work was done partly in Oslo, partly in the Mineralogical Laboratory in Cambridge. The arrangement of the lead ions was readily found with considerable accuracy; the other ions, however, being very light as compared with the lead ions, contribute so little to the intensities in the photographs that no definite conclusions as to their positions can be drawn from the X-ray data alone. Because of this difficulty, which does not seem surmountable (except perhaps by means of accurate quantitative intensity measurements), I did not publish the results I had arrived at. Some years later E. Onorato published a paper on the structure of phosgenite[1]. Although his determination of the lead positions is in close agreement with my own it was obvious to me that the structure could not be correct in other respects. As will be seen below, the unit cell used by Onorato ($a = 8.1$Å, $c = 4.4$Å, containing 2 $Pb_2Cl_2CO_8$) is only a pseudo cell, the real unit cell being twice as large (at least). In addition the structure given by Onorato shows the highly improbable feature of containing no CO_8-groups, which are otherwise characteristic of all known carbonate structures. For the above mentioned reason I am not able to prove the existence of CO_8-groups in the phosgenite structure, but on the other hand I find it equally impossible to prove the opposite from X-ray data. Since the appearance of Onoratos,

---

[1] E. Onorato, La struttura della Fosgenite. Periodico Min. Roma, *5*, 37, 1934. Abstract in the Strukturbericht of the Z. Krist., Ergänzungsbd. *3*, 1937.

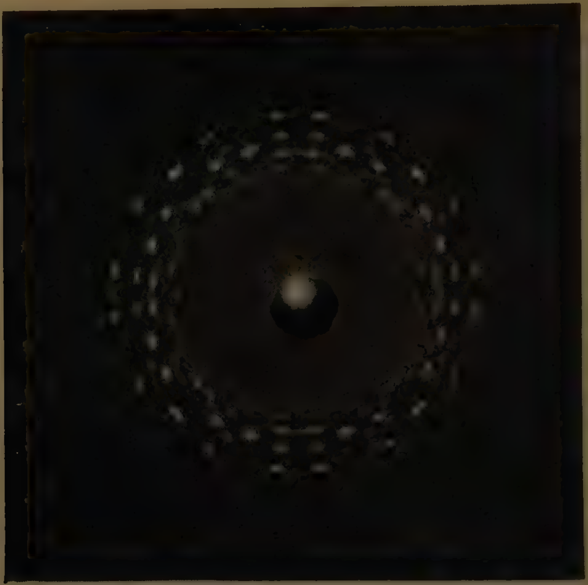

Fig. 1. Laue photograph of phosgenite, X-ray beam nearly perpendicular to 001.

paper no one seems to have published anything on the structure of phosgenite. Using my X-ray data and known ionic radii &c. I am able to propose a structure which is at least in some respects more correct and on the whole seems much more reasonable.

The complete ditetragonal symmetry of the Laue photographs of phosgenite is sufficiently known from earlier work; it is clearly demonstrated by the figures 1 and 2.

The powder photographs, which contain more than 50 different α-lines, lead to the above mentioned (pseudo) cell and may be completely interpreted in terms of this cell. Fig. 4 shows the indices of most of these lines and their observed intensities, by the full line in the diagram. By the powder method were also examined the synthetic compounds $Pb_2Cl_2CO_3$ and $Pb_2Br_2CO_3$. These compounds were obtained by a method described by A. de Schulten[1]. The resulting preparations consisted of beautiful tabular tetragonal crystals about 0.2 mm across. As remarked by de Schulten, the Cl-compound agrees

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[1] A. de Schulten, Sur la production artificielle à la température ordinaire de la phosgénite et de la phosgénite bromée. Bull. Soc. fr. de Min. *20*, 191, 1897. Abstr.: Z. Krist. *31*, 75, 1899.

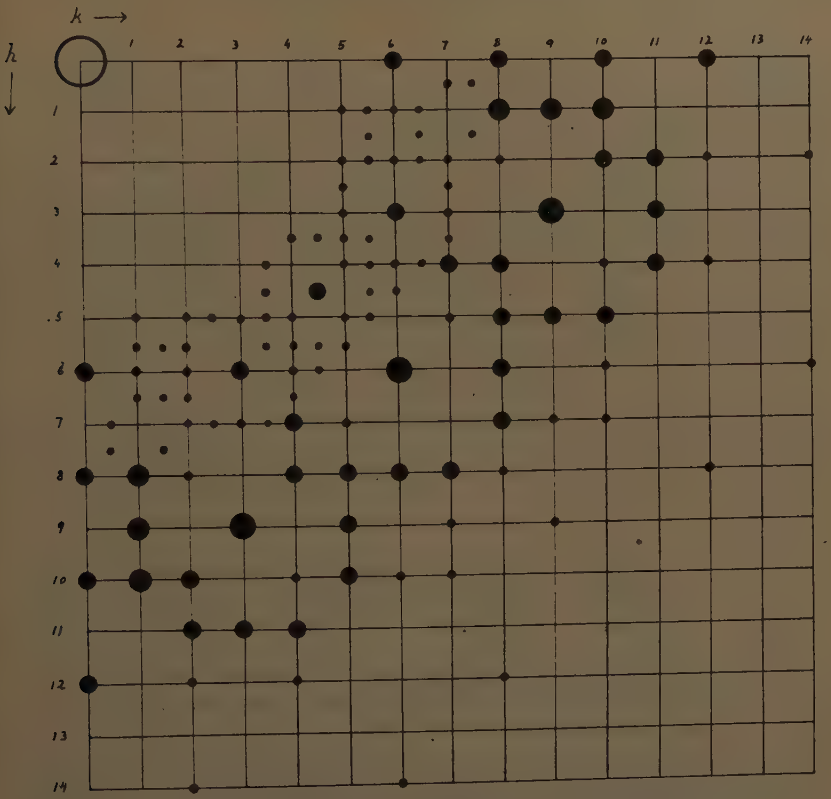

Fig. 2. Gnomonic projection of one quadrant of the Laue photograph Fig. 1.

completely with phosgenite in all its properties, and the Br-compound is very similar. The latter shows, naturally, the higher density. Optical examination shows that the Cl-compound is uniaxial positive, while the Br-compound is uniaxial negative; the birefringence is rather low in both cases. The powder photographs of natural and synthetic phosgenite are identical in all details; those of Br-phosgenite are similar apart from slight differences in spacings and intensities. It is obvious that these substances belong to the same structure type. A Laue photograph of one of the Br-phosgenite crystals, which was very faint on account of the small size of the crystal, was similar to that of natural phosgenite in symmetry and positions of the main spots (Fig. 3). It may be remarked that the intensity differences between the powder photographs of phosgenite and Br-phosgenite

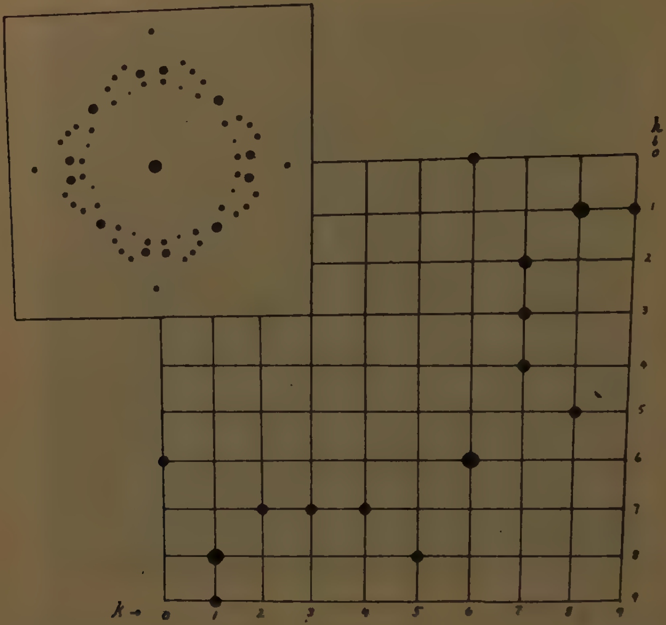

Fig. 3. Br-phosgenite. Laue photograph and gnomonic projection. (The photograph is too faint for direct reproduction.)

are probably due to the different scattering powers of Cl and Br. It might be thought that this would help to determine the positions of the halogene ions in the structure, but unfortunately the intensity differences are too insignificant to be of any practical use. — By the method of de Schulten I also obtained similar crystals of a compound $Pb_2(Br,Cl)_2CO_3$. Although I started from a solution containing equivalent quantities of Cl and Br, the resulting mixed crystals were obviously enriched in Br, as shown by the lattice constants given below. The mixed crystals were anomally biaxial, their central parts being probably especially rich in Br.

The following quadratic forms explain — for $Cu_{K\alpha}$-radiation — all the lines observed in the powder photographs.

Phosgenite (natural and
synthetic): $\sin^2\varphi = 0.008975\ (h^2 + k^2) + 0.03015\ l^2$.
Br,Cl-phosgenite: $\sin^2\varphi = 0.00860\ (h^2 + k^2) + 0.0290\ l^2$.
Br-phosgenite: $\sin^2\varphi = 0.00850\ (h^2 + k^2) + 0.0287\ l^2$.

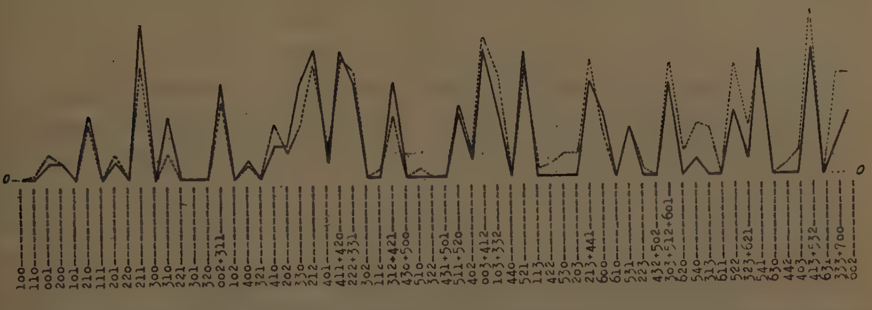

Fig. 4. Observed (full line) and calculated (dotted line) intensities for powder photographs of phosgenite, the calculated intensities being obtained from the Pb-lattice alone.

The coefficients have been fixed by means of rocksalt as a gauging substance and are thus fairly accurate. They give the following lattice dimensions.

	a	*c*	*c/a*	*d*
Phosgenite	8.125 Å	4.430 Å	0.5453	6.19
(Br, Cl)-phosgenite	8.299 Å	4.519 Å	0.5445	
Br-phosgenite	8.347 Å	4.541 Å	0.5441	6.62

These figures are uncertain only by a few units in the last place. *d* is the calculated density for $Z=2$. de Schulten (l. c.) determined directly the densities of his synthetic crystals and found 6.134 for Cl-phosgenite and 6.550 for Br-phosgenite. (The agreement is very good, but, as in many other cases, the directly measured densities are systematically a little lower than those calculated from X-ray measurements, possibly because of invisible cracks or cavities in the crystals.)

The interpretation of the powder photographs leads uniquely to the following arrangement of the Pb ions, in terms of the (pseudo) cells given above.

$uu0$; $\overline{u}\overline{u}0$; $u+\frac{1}{2},\frac{1}{2}-u,0$; $\frac{1}{2}-u,u+\frac{1}{2},0$, u being close to 1/6. This Pb-lattice gives calculated intensities which on the whole agree well with the observed ones (Fig. 4), showing, as might be foreseen, that the other ions contribute relatively little to the intensities. Taking into account the halogene ions, placed in positions

$vv\frac{1}{2}$; $\overline{v}\overline{v}\frac{1}{2}$; $v+\frac{1}{2},\frac{1}{2}-v,\frac{1}{2}$; $\frac{1}{2}-v,v+\frac{1}{2},\frac{1}{2}$ (forming layers midway between the basal Pb-layers) and with a suitable value for v, the the agreement is considerably improved; however, we now proceed to consider the oscillation photographs, which yield more complete data. The oscillation photographs were taken with a Bernal X-ray photogoniometer[1], using a copper anticathode. Axes of rotation were the crystallographic axes *c* and *a*. Only natural phosgenite could be examined by this method. These photographs give the following dimensions of the unit cell: *c* about 8.8Å, *a* about 8.1Å. The *c*-axis used above has thus to be doubled, and more accurately the cell dimensions come out:

$$c = 8.860\ \text{Å} \qquad a = 8.125\ \text{Å} \qquad c/a = 1.091.$$

This cell naturally contains 4 $Pb_2Cl_2CO_3$. Its axial ratio agrees closely with the value given for phosgenite in mineralogical textbooks, which is 1.0876. — The interpretation of the oscillation photographs was done by the graphical method worked out by J. D. Bernal[2]. The result, with observed intensities, is shown in Table 1. It is seen at once that the pseudo cell is highly pronounced, the indices with l odd being represented by relatively very weak spots only. This is a consequence of the Pb-arrangement, which can — at least very approximately — be described in terms of the pseudo cell as above. Of abnormal spacings we have the following: $h0l$ is halved if h is odd. This is also seen from the Laue photograph, fig. 2. This applies to the space groups D_{2d}^7 — $P\bar{4}b2$, C_{4v}^2 — P4bm, D_{4h}^5 — P4/mbm, which are also in accordance with the Laue symmetry of phosgenite. If phosgenite is piezoelectric, as stated by Onorato (l. c.), this is in favour of the two former space groups, which belong to the tetragonal scalenohedral and the ditetragonal pyramidal class respectively. From observations of the crystal faces the holohedral and the trapezohedral classes have been proposed. (Onorato arrived at the trapezohedral space group D_4^2 — $P42_1$, but this is not in accordance with my X-ray data, even when the above pseudo cell is used.) The question as to the correct space group can hardly be decided with certainty. In view of the relatively very few and weak observed spots with index l odd also the space groups with no abnormal spacings (D_{2d}^1, D_{2d}^5, C_{4v}^1, D_4^1, D_{4h}^1) — and even still larger unit cells — strictly have to be considered. However, as the observed distribution of intensities definitely points to the above three space groups we may assume

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[1] J. D. Bernal, J. Sc. Instr. *4*, No. 9, 1927: *5*, Nos. 8/9, 1928: *6*, Nos. 10 and 11, 1929.

[2] J. D. Bernal, On the Interpretation of X-ray, Single Crystal, Rotation Photographs. Proc. Roy. Soc. (A) *113*, 117, 1927.

Table 1.

Estimated intensities from oscillation photographs.

hkl	Int.	*hkl*	Int.	*hkl*	Int.	*hkl*	Int.	*hkl*	Int.
100	-	101	-	102	-	103	-	104	-
110	8	111	-	112	10	113	-	114	?
200	6	201	-	202	8	203	4	204	15
210	30	211	-	212	30	213	6	214	20
220	8	221	-	222	15	223	4	224	6
300	-	301	-	302	-	303	-	304	-
310	30	311	-	312	20	313	-	314	20
320	2	321	-	322	2	323	2	324	-
330	30	331	6	332	30	333	4	334	30
400	15	401	-	402	15(?)	403	2(?)	404	15(?)
410	30	411	-	412	30(?)	413	6	414	20
420	15	421	4	422	20	423	-	424	15
430	6	431	-	432	2	433	-	434	-
440	20	441	-	442	10	443	2	444	10
500	-	501	-	502	-	503	-	504	-
510	2	511	-	512	2	513	2(?)	514	6
520	30	521	-	522	30	523	6	524	20
530	15	531	-	532	25	533	6	534	20
540	30	541	-	542	30	543	6	544	20
550	2	551	-	552	?	553	-	554	4
600	25(?)	601	-	602	20(?)	603	-	604	30(?)
610	2(?)	611	-	612	2	613	-	614	?
620	15	621	-	622	15	623	2	624	15
630	?	631	-	632	-	633	-	634	-
640	20	641	2	642	15	643	-	644	15
650	6	651	-	652	-	653	2	654	-
660	30	661	-	662	30	663	4(?)	664	30
700	?	701	-	702	?	703	-	704	?
710	2	711	-	712	?	713	-	714	4(?)
720	30	721	-	722	25	723	2	724	20
730	25	731	4	732	20	733	-	734	20
740	25	741	-	742	25	743	4	744	20
750	6	751	-	752	6	753	?	754	10
760	4	761	-	762	4	763	-		
770	6								
800	15	801	-	802	15	803	6	804	15
810	30	811	-	812	30	813	4	814	20
820	4	821	-	822	8	823	4	824	15
830	?	831	-	832	-	833	2	834	-
840	10	841	-	842	15	843	4	844	15
850	30(?)	851	-	852	30(?)	853	8		
860	15	861	-	862	15	863	-		
900	?	901	-	902	?	903	-	904	?
910	30	911	-	912	20(?)	913	2		
920	2	921	-	922	-	923	4		
930	30	931	4	932	30	933	6		
940	2	941	-	942	6				
950	30								
001	-	002	30	003	1	004	30	005	6

that the structure is built on one of these. Further, as the point coordinates in the space group P4/mbm may be built up of suitable sets of point coordinates in each of the space groups P4bm and P$\bar{4}$b2, the structure may probably be approximately described by means of the former (holohedral) space group, even if the correct space group should be one of the latter ones. We therefore try to build the structure on the space group P4/mbm; if necessary we can afterwards displace the points in accordance with P4bm or P$\bar{4}$b2.

The Pb-arrangement found from the powder photographs is now expressed as follows.

8Pb in (k): $u, \frac{1}{2}-u, v;\ u+\frac{1}{2}, u, v;\ \frac{1}{2}-u, \bar{u}, v;\ \bar{u}, u+\frac{1}{2}, v;$
$\bar{u}, u+\frac{1}{2}, \bar{v};\ \frac{1}{2}-u, \bar{u}, \bar{v};\ u+\frac{1}{2}, u, \bar{v};\ u, \frac{1}{2}-u, \bar{v},$
u being close to 1/6 and $v=1/4$.

Naturally this Pb-lattice alone gives calculated intensities which agree roughly with the observes ones, but it does not explain the spots for which index l is odd. Such spots would occur if v had a value differing a little from 1/4; this possibility will be discussed later; in any case the deviation from 1/4 must be exceedingly small, as otherwise some of the spots with index l odd would have to be much stronger than they actually are. We therefore for the present assume that v is exactly 1/4. As mentioned above the agreement between calculated and observed intensities is improved by taking into account the Cl-ions placed in layers midway between the basal Pb-layers, that is 4Cl in (g) and 4Cl in (h), with suitable u-values. However, a consideration of this Pb-Cl-arrangement shows that it hardly leaves spaces where the C and O can be placed; moreover it does not necessarily explain the spots with index l odd. Now it is possible to exchange one of the Cl-layers, e. g. (h), against a layer of CO_3-groups; the 4 CO_3-groups can be placed lying wholly in one plane and still in full accordance with the known ionic distances. This arrangement is possible in the real unit cell only, and thus helps to explain the occurrence of spots with index l odd. It still remains to put 4 Cl into the structure. These may be placed in the centres of the Pb-squares in the Pb-layers themselves (positions (e) with $u=1/4$). With suitable parameter values the whole arrangement can be brought in accordance with the known ionic radii. The structure is then expressed as follows, including the above Pb-positions.

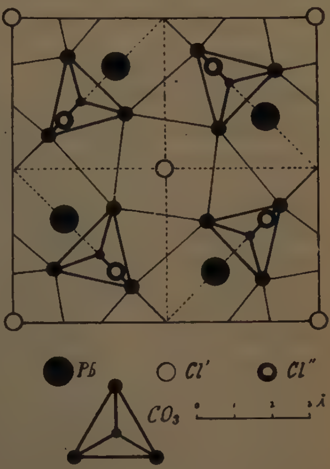


Fig. 5. Proposed structure of phosgenite. A unit cell projected on the basal plane. 8 Pb in heights 1/4 c and 3/4 c; 4 Cl' in heights 1/4 c and 3/4 c; 4 Cl'' in height 0; 4 CO_3 in height $\frac{1}{2}c$. Thin full lines connect neighbouring O-ions.

4 Cl in (g): $u, u+\frac{1}{2}, 0; \frac{1}{2}-u, u, 0; u+\frac{1}{2}, \bar{u}, 0; \bar{u}, \frac{1}{2}-u, 0,\ u \sim 1/6.$
4 Cl in (e): $00u;\ 00\bar{u};\ \frac{1}{2}\frac{1}{2}\bar{u};\ \frac{1}{2}\frac{1}{2}u,\ u = 1/4.$
4 C in (h): $u, u+\frac{1}{2}, \frac{1}{2}; \frac{1}{2}-u, u, \frac{1}{2}; u+\frac{1}{2}, \bar{u}, \frac{1}{2}; \bar{u}, \frac{1}{2}-u, \frac{1}{2},\ u \overset{=}{<} 1/4.$
4 O in (h) with $u \sim 1/8.$
8 O in (j): $uv\frac{1}{2};\ v\bar{u}\frac{1}{2};\ v+\frac{1}{2}, u+\frac{1}{2}, \frac{1}{2}; u+\frac{1}{2}, \frac{1}{2}-v, \frac{1}{2};\ \overline{uv}\frac{1}{2};\ \bar{v}u\frac{1}{2};\ \frac{1}{2}-v, \frac{1}{2}-u, \frac{1}{2}; \frac{1}{2}-u, v+\frac{1}{2}, \frac{1}{2},\ u \overset{=}{>} 1/8,\ v \sim 1/6.$

Fig. 5 shows a unit cell of this structure, projected on 001. The CO_3-groups have been assumed to have the usual shape of equilateral triangles with side lengths about 2.2Å. They form a plane sheet, the shortest distance O-O between neighbouring CO_3-groups being about 2.6Å, which is the value to be expected from the ionic radius of O^{2-}; each O has 3 neighbours in this distance, but as to the arrangement of these there are 2 kinds of O, corresponding to the above division of the O-positions into 4+8 equivalent positions. This CO_3-sheet looks very plausible. If it really occurs in the phosgenite structure we should expect the phosgenite to be strongly optically negative, while it is actually weakly positive. It is possible, therefore, that the planes of the CO_3-groups themselves are inclined or even perpendicular to the plane of the sheet. Perpendicular positions can be effected in the same space group, but it seems difficult then to arrange for sufficient distances to the Pb and Cl ions. Inclined positions are possible in the two other space groups, P4bm and P$\bar{4}$b2, which have no center of symmetry. However, the optical character of phosgenite may be due to other things than the CO_3-groups; as mentioned the substitution of Br for Cl in phosgenite changes the optical sign. Thus speculations as to the exact positions of the CO_3-groups are obviously futile. — Each Pb is surrounded by 5 Cl in the distances 3.0 Å, 2.9Å, 3.3Å (calculated from the ionic radii: 3.13Å) and 2 O distant 2.5Å (calculated 2.64Å), further 3 O about 3.1Å distant. The shortest distance Cl-Cl is about 3.7Å (calculated 3.62Å) and Cl-O about 2.8Å (calculated 3.13Å). Thus the structure is in accordance with the ionic radii involved. As emphasized above it is not possible to check this arrangement of the lighter ions by means of the X-ray data, the fixed Pb-arrangement in combination with any arrangement of the other ions giving calculated intensities which agree fairly well with the observed intensities. This difficulty is further increased by the following fact. It is seen from the intensity table that there is a distinct increase of intensities in the series 001, 003, 005, and on the whole that the spots *hk*3 are on the average much stronger than *hk* 1. Such a feature points to small deviations in the *c*-direction from the positions given by the pseudo cell (half cell), and the rather high intensity of 005 makes it probable that also the Pb ions are involved in these deviations. This would mean that the parameter v in the Pb-positions is not exactly 1/4; in fact a value about 0.24 or 0.26 gives good agreement for the basal reflections. This small uncertainty in the Pb-positions naturally makes it still more difficult to trace the intensity contributions of the lighter ions.

Summary.

Natural and synthetic phosgenite, $Pb_2Cl_2CO_3$, as well as the synthetic compound $Pb_2Br_2CO_3$ (Br-phosgenite) and a mixed crystal of these two components, belong to the same tetragonal structure type. The lattice constants and calculated densities are as follows.

	c	*a*	*c/a*	*d*
Phosgenite	8.860 Å	8.125 Å	1.091	6.19
Br-phosgenite	9.082 Å	8.347 Å	1.088	6.62

For the mixed crystal were found values intermediate between these. The unit cells contain 4 "molecules"; this is in full agreement with the observed densities. The following arrangement of the ions in the phosgenite cell is proposed. See Fig. 5.

Space group: P4/mbm. Coordinate notation after Wyckoff:

8 Pb in (k) with $u \sim 1/6$ and $v \sim 1/4$.
4 Cl in (g) with $u \sim 1/6$.
4 Cl in (e) with $u = 1/4$.
4 C in (h) with $u \overset{=}{<} 1/4$.
4 O in (h) with $u \sim 1/8$.
8 O in(j) with $u \overset{=}{>} 1/8$ and $v \sim 1/6$.

The Pb-positions have been determined from the X-ray data, the positions of the other ions mainly from considerations of the ionic radii. The above space group has not been proved to be the correct one; P4bm and $P\bar{4}b2$ are equally possible, or perhaps even more probable as they would explain the piezoelectricity which is said to have been observed in phosgenite. The above structure is a kind of ideal arrangement, especially as to the position of the CO_3-groups. In the real structure these may be inclined to the basal plane.

Oslo, Mineralogisk-geologisk museum,
November 1944.

Printed February 26th, 1945.

KONSERVATOR ANDERS NUMMEDAL

Minnetale i Norsk geologisk forening 23. mars 1944

AV

HALVOR ROSENDAHL

Konservator Nummedal døydde 7. mars 1944, 77 år gamal, etter å ha vore sjuk i nokon tid. Frua hans døydde nokre viker før, og det gjorde vel sitt til, at det tok fortare slutt med hans liv ogso. Han hadde vore medlem av Norsk geologisk forening frå 1912. Siste gong var han her i foreninga på møtet 17. april 1941, då han viste fram eit fund av dansk stein på Tjøme. So lenge han hadde helsa, var han ofte på møta i tida etter 1922, då han kom til Oslo. I denna tida har han halde 6 foredrag i Norsk geologisk forening:

5/4 1923. Om flintplassene
8/4 1926. Steinaldersfundene i Alta
9/2 1928. Beretning om en reise i Østfinnmark
26/9 1929. Et steinaldersfund i Ski
2/2 1933. Kvartærgeologiske spørsmål i Østfold og Møre
17/4 1941. Et fund av dansk stein på Tjøme

27/9 1931 leida han ein ekskursjon til steinaldersbuplassen på Sørli i Ski.

Anders Jonsen Nummedal er fødd 27. januar 1867 i Vik i Sogn. Etter gamal skikk kalla han seg med farsnamnet etter faren Jon Olsen Nummedal. Han kom inn på eit artiumkursus og vart student på Voss 1889. Seinare fekk han ymse skulearbeid i Oslo og arbeidde samtidig med studiet til embetseksamen, som han tok i 1900. Same år vart han adjunkt, seinare lektor, i Kristiansund, der han bleiv til 1922, då han vart knytt til Universitetets oldsaksamling i Oslo.

Medan han studerte i Oslo, var han mykje med på ekskursjonar og vart ein god kjennar av kvartærfossila. Mange er dei fund, stud. real. Nummedal gjorde, og som no ligg i samlinga på Geologisk museum. I Brøggers og Øyens avhandlingar frå den tida står Nummedals navn fleire stader nemnt. Alt no synte seg den eigne evna til å finna seg til rette i naturen, den som seinare skulde føra til so store resultat.

I Kristiansund tok han til med kvartærgeologisk gransking. Både skulebarna og andre tykte ofte, han var ein raring, som gjekk slik ute i all slags ver og grov.

På ein av desse turane sine fann han i slutten av oktober 1909 nokre flintstykke, som såg ut til å vera handsama av menneske. Han sende deim til oldsaksamlinga åt vitenskapselskapet i Trondheim og fekk svar frå styraren, K. Rygh, at det var oldsaker, og at det var von om å finna fleir på same staden. Han gjekk atter til finnestaden og fann på kort tid 50 flintstykke, som alle var meir eller minder arbeidt og emna til av menneske, skiveøksar, knivar, pilespissar og borar. Ogso desse flintstykka vart sende til Rygh, og han kunde no slå fast, at Nummedal hadde funne ein verkstad eller kanskje ein bustad frå steinalderen. Fosna-kulturen var funnen.

I Oldsaksamlingas aukeliste for 1910 kom dei nye funda først på prent. Det var då innkome 8000 flintstykke frå ytre Nordmøre. Rygh gav deim ein serskilt omtale. Slike fund var kjende før men ikkje på lang leid med so stort materiale. Rygh sette namnet flintplassar på fundstadene.

I si første avhandling om flintplassane, Oldtiden 2, 1912, skreiv Rygh, at Nummedals oppdaging var ein av dei vigtigaste hendingar i norsk arkæologi i dei seinare åra. Meir enn 100 flintplassar var då kjende. Sikkert var dei arbeidsplassar, nokre var sikkert ikkje buplassar, men andre kunde vel ogso ha vore buplassar. Skiveøksane av flint overtyda Rygh om, at dei var frå skjeldungetida i den eldre del av den neolithiske tida, endå andre reidskapar hadde eldre bragd.

Det har seinare vore ulike meiningar om flintplassane; men Rygh har visseleg på første hand i hovudsaka råka den rette alderstydinga. Truleg er vel det eldste av flintplasskulturen noko eldre og kan førast attende inn i borealtida, Ancylus-tida. Det hevda Nummedal ogso i 1912.

Det, dei kjende den gongen av norske steinaldersfund, vart sett i desse gruppene:

Yngre steinalder, som omfata dei vel kjende yngre typer i kronologisk system.

Arktisk steinalder, som ikkje var noko reint kronologisk omgrip systematisk likeverdig med yngre steinalder. Til arktisk steinalder var rekna ei rekke med steinalders fundstader langs kysten frå Trøndelagen til Aust-Finnmark, rester etter eit veidefolk utan høg reidskapskultur. Dei brukte mykje skifer; men ogso flint og annan stein vart nytta. Skjeldungen på Steinkjer var rekna med til arktisk steinalder og like eins nokre fundstader sørover vestlandskysten, soleis Vespestad på Bømlo og Holeheia på Jæren, som var den synste.

Viste-dungen på Jæren vart ikkje rekna til arktisk steinalder men til skjeldungekulturen, svarande til den danske Kjøkkenmøddingkultur.

Nøstveit-kulturen var ogso ein primitiv reidskapskultur, serleg utbreidd over søraustlandet men ogso vestover til Lista, ja kanskje nordover vestlandskysten, der det ikkje er lett å skilja han frå arktisk steinalder. Nøstveit-folka var ogso veidefolk, berre med dei mest livsturvande reidskapar av dei bergartar, dei lettast fann på staden. Den vanlege meining var, at dei var samtidig med skjeldungefolket, eldre enn arktisk steinalder.

I motsetnad til yngre steinalder og arktisk steinalder vart skjeldungekulturen og Nøstveit-kulturen kalla eldre steinalder, d.v.s. eldre nordisk steinalder. Men dermed var det ikkje sagt, at dei var palæolithiske i europeisk meining.

I eit storfelt vidsyn hadde Andr. M. Hansen tolka vår fornalders utvikling:

Den gamle veidekulturen, som i Danmark var bore oppe av Kjøkkenmøddingfolket, kom seinare til Noreg. Då var alt det neolithiske folket, som ein etter dei store steingravane ogso kallar megalithfolket, kome til Danmark med jordbruk og husdyr. Veidefolket hadde av husdyr berre hunden; det var kome frå aust eller

søraust. Megalithfolket kom sunnanfrå. Ein kan fylgja megalithkulturen attende til Midhavs-landa, der det i den tida var bronsealder, Egypt med steinpyramidane, og dei gamle kulturland i Asia, so ogso denne kulturstraumen stort set gjekk frå søraust til nordvest. Til Noreg kom megalithfolket først langt oppe i yngre steinalder, ja etter Hansens meining først i nordisk bronsealder i tida for vinnamna. Det er Andr. M. Hansens ariske landnåm; veidefolket var etter hans meining anarisk. Men veidekulturen var ikkje slutt for det, om det nye folket var kome. Veidekulturen heldt fram i Noreg, først i Nøstveit-kulturen og Viste-kulturen, seinare i den arktiske steinalder. Den eldre steinalders og den yngre steinalders folk er soleis ikkje skilde frå kvarandre i tid men berre i stad og levevis. Ein kan derfor av reidskapstypane ikkje slutta noko visst om tida. Dei har stende i samband med kvarandre, det ein kunde kalla kultursymbiose, på same vis som i våre dagar innlandseskimoar og sjøeskimoar, fjellfinnar og sjøfinnar. Det såg Andr. M. Hansen før nokon annan, og nyare granskingar i Danmark har vist, at han såg rett.

Men når han meinte, at det sørskandinaviske veidefolket var anarisk, har han urett. Maglemose- og Ertebølle-folket, Kjøkkenmøddingfolket, ser ut til å ha vore like ariske som megalithfolket. Truleg har han ogso urett i det, at det var den gamle veidekulturen, som gjenom Nøstveit- og Viste-kulturen heldt fram i den arktiske steinalder. Nummedal var og usamd med honom i det. Men hans hypotese var logisk oppbygd, for det arktiske steinaldersfolket var ogso anarisk, meinte han. At det i den arktiske steinalderskrinsen inngår anariske folk, er rett nok, og at det i Noreg har vore finnar, kan ein trygt gå ut fra. Om det lengre sør i det arktiske steinaldersområdet har vore andre, eventuelt ariske, folk, veit ein ikkje. Etter Hansen skulde det arktiske steinaldersfolket ikkje ha vore finnar i vår meining av ordet, men eit anna anarisk folk, som han kalla finnar eller skridfinnar; til deim rekna han ogso dei folka, dei i gamle dagar kalla kvænar og bjarmar. Dette gamle arktiske veidefolket er no mest oppblanda med den yngre ariske delen av det norske folket, men lever endå att i sjøfinnane. Fjellfinnane derimot skal ikkje vera ætta frå dei gamle hypotetiske skridfinnane, men er komne seinare. Kor mykje rett, det er i Hansens hypotese, er no ikkje godt å segja; for endå har me ikkje grunnlagsmateriale nok til å avgjera det. Til den arktiske kulturkrinsen i våre dagar, som ein godt kunde kalla den resente arktiske steinalderen, høyrer dei kjende anariske folka rundt Nordishavet, finnar, samojedar, tungusar, tsjuktsjar, eskimoar. Slik var det vel ogso i den gamle arktiske steinalderen. Då var ogso ein stor del av Skandinavia med i det sirkumpolære arktiske kulturområdet. Arktisk steinalder er derfor eit sers godt namn på denne kulturen, i fullt like stor mun no som i 1871, då Oluf Rygh innførde namnet.

Nummedals oppdaging viste, at det so langt nord som på Møre hadde vore folk i eldre nordisk steinalder, i mykje eldre tid enn ein før visste. Og det var ikkje småting, Nummedal grov fram, det var flintstykke i titusental. Alle stader, han grov, fann han flintplassar, frå Sotra til Nord-Trøndelag. Den store bruk av flint heng saman med, at det då endå var mykje flint å finna i det kvartære materialet. At flinten er funnen på staden og ikkje innførd, er det ingen tvil om.

Nummedal har ogso granska hellerer og holer. Den første avhandlinga hans var om Dalehelleren og Valseshola ved Kristiansund. Det viser seg, at mesta alle hellerer og holer, som låg tilgjengeleg, har vore nytta. Veidefolk har hatt deim til millombils bustad i ymse tider, frå steinalder til jernalder og heilt opp i historisk tid. Det er ogso truleg, at fredlause har hatt tilhald der, og dei har vel ogso vore brukt til bergingsrom i ufredstider. Det er aldri funne bustader frå eldre steinalder i holer i Noreg. Nokon egte holekultur med holemenneske har me ikkje hjå oss. Den første større gransking av ein heller i Noreg vart gjord av H. Reusch i Sjonghelleren på Valderøy i Sunnmøre 1875.

Mest kjend er vel Nummedal for sine granskingar av steinalderen i Finnmark. Det var Institutet for samanliknande kulturforsking, som tok til med dette arbeidet i 1925, og Nummedal vart utsedd til å setja det i verk. Den første sumaren 1925 var han i Alta og kom her midt opp i ein gamal steinalderskultur, der det ogso hadde vore nytta norsk flint. 8. april 1926 heldt han foredrag i Norsk geologisk forening om steinaldersfunda i Alta, og den første prenta publikasjonen kom i Norsk geologisk tidsskrift 9, 1926. Han innførde her namnet Komsa-kulturen etter fjellnamnet Komsa i Alta, og samtidig sette han namnet Fosna-kulturen på flintplasskulturen på Møre. Han sette desse kulturane nært kvarandre. Dermed skulde alderstydinga for Fosna-kulturen, skjeldungetid og noko attende i borealtida, ogso gjelda for Komsa-kulturen. Etter alt, ein no veit, er det rett.

13 sumrar var han i Finnmark og grov fram ei mengd med steinalderssaker, frå Alta til Peisen. Materialet er flint og gode kvartsitar frå den eokambriske formasjon. Det er mest avfall og dårlege reidskapar, men det er gode skiveøksar i bland. Liksom ved flintplassane var slike fund kjende før, men ikkje på lang leid i so overhendig stor mengd.

I 1929 fann han liknande ting i Austfold. Dei låg mykje høgare enn Nøstveit-funda, over 160 m o. h., millom Litorina- og Pholas-nivået. Skulde dei i tid svara til høgda over havet, som Andr. M. Hansen meinte for Nøstveit-funda, måtte dei vera svært gamle, langt tilbake i den boreale tida, Ancylus-tida. Dei øvste funda i Finnmark låg i tilsvarande høgd, og alt i 1912 hadde Nummedal hevda, at flintplassane var frå Ancylus-tid.

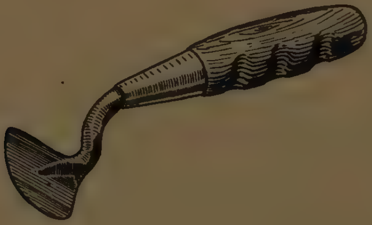
Det spørsmålet melder seg no, om desse tri fundgruppene høyrer til same kulturområde. Finnmark har utan tvil høyrt til det arktiske område. Truleg må vel ogso Møre reknast til det arktiske område i den tida. Om Austfold ogso høyrde til det arktiske kulturområde, er eit ope spørsmål. Men slike systematiseringsspørsmål har ofte mindre realitet i livsrøynda enn i våre spekulasjonar.

I dei siste åra, Nummedal var i Finnmark, 1935—1937, arbeidde han mest med yngre steinalder inst i Varangen. I Karlebotn grov han i tuftene etter ein steinaldersby med 72 talde hus. Keramikken, dei fann der, hadde tydeleg samband med typer lengre aust. Likevel vilde han gjerne ha det til, at den yngre steinalderen i Finnmark stod i samband med yngre sørskandinavisk steinalder, megalithikum. Men det er visst ikkje rett. Den yngre steinalderskulturen i Finnmark er arktisk, sameleis som ogso Komsa-kulturen er. Frå dei eldste Komsa-fund til Karlebotn-byen er det gjenge tusener av år med både etniske og kulturelle omskifte, men alt innanfor den arktiske kulturkrinsen,

Eg har i dette oversynet over Nummedals arbeid nemnt Andr. M. Hansen. Når eg no til slutt vil segja nokre ord om Nummedals person, fell det ogso naturleg å samanstella desse tvo. Opphavleg var dei begge geologar, og først seinare kom dei inn i arkeologien. Med dei ulike naturane deira verka det ulikt på deim. Hansens skapande fantasi vart stimulert av å gå frå utsyn til utsyn, og han bygde med si sjølvmedvitande konstruktive evne opp meir, enn han strengt teke hadde grunnlagsmateriale til. Han hadde sin styrke ved skrivebordet. Nummedal var sterkast ute i naturen, og han hadde materiale-tilfang til meir enn han bygde opp. I arkeologien kjende Nummedal seg autodidakt utanfor fagkrinsen, og det førde med seg ei overvurdering av det, som var innanfor, og ein tilsvarande skort på tiltru til eigen autoritet. Det kan forklåra den uvissa, han hadde i kritisk vurdering og tolking. Han bygde meir på andre autoritetar enn på si eiga dømekraft, og denne noko ukritiske ovtrua på autoriteten kunde føra honom på villstrå. Det var ei tid, då han vilde føra Fosna- og Komsa-kulturen attende til palæolithisk tid, Aurignacien, ja ogso Moustérien har vore nemnt. Grunnen til det var frå førsten ingen annan enn den, at han i bøker om palæolithiske kulturar i Millomeuropa fann bilete av steinreidskapar, som likna dei, han fann her heime. Seinare voks trua på det fast i honom til noko med hans eige livsinnhald, og han prøvde ogso å finna geologisk studnad for det. Soleis har han i avhandlinga om flintplassane i 1922 ei utgreiding om dei geologiske tilhøva i Sunndalen. Han meinte, at det fossilførande leiret på Hoås, der han ogso fann Portlandia arctica, var interglasialt, og at Hoås-terrassen, som ligg over det, var den ytste morenen frå siste istid. Utanfor Hoås hadde det soleis vore isfritt i siste istida, og det skulde gjeva auka vilkår for, at menneska hadde kunna bu der i seinglasial tid, ja kanskje i sjølve glasialtida og interglasialtida. Ja han var verkeleg inne på den tanken, at dei kjende fornaldersfunda i Noreg kunde skriva seg fra glasiale eller interglasiale menneske, og i dette fekk han til skade for seg sjølv studnad frå ymse leider her i landet. Men heilt viss på det kjende han seg ikkje; for ein slik hypotese, som stod i strid med den endeframme utsegna av funda ute i naturen, førde til sjølvmotsegjing og inkonsekvens. Dei instinkta, som ute i naturen førde honom på rett veg, var ikkje med honom innanfor arbeidsromets fire vegger.

Nummedal hadde dei eigenskapar, som karakteriserer det noble menneske, og som alltid har vore agta høgt i Noreg. Han var av reint norsk opphav og hadde mykje av nordmannens lite formelle evne og åtferd, og han vilde gjerne halda påstendig på si meining. Det kvite håret dekte ein hovudskalle av rein nordisk type, dei blå augo hadde ein vakker og usegjeleg venleg dåm. Men når nokon sa honom imot i slike ting, som for honom var vigtige, då lyste dei av motstand, og han var ikkje av dei, som gav seg. Norsk kultur og norsk mål låg honom nært på hjarta, og det var hugnad for honom, at han i si verksemd med å læra oss vår fornalder å kjenna, kunde vera med i dette kulturarbeid, som vår tid treng so hardt.

Til sine utgravingar brukte han ein murarskei av sterkt stål. I 1927 var den første reidskapen hans so langt nedsliten, at det berre var ein liten runding att av stålplata, og i treskaftet var det forer etter fingrane. Han var då alt på veg med å slita ned ein ny. Om eg ikke sjølv hadde set det, vilde eg ikkje ha trutt, at sovore var mogleg. Denne nedslitne spaden vil eg setja til symbol på den trottuge, trugne granskaren med den indre elden, ikkje ein logebrennande eld, men ein intens glod-eld, som endåtil ikkje stålet kunde stå imot, og som berre dauden kunde sløkkja.



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THE THERMODYNAMICS OF THE EARTH'S CRUST

I

Preliminary Survey of the Principal Forces and Reactions in the solid Crust.

BY

HANS RAMBERG

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Introduction.

During field studies and theoretical considerations of metamorphism and metasomatism I have arrived at the opinion that the apparently so multifarious and intricate processes of metasomatism are led by very simple principles. By recognizing these principles, we may get a clear and simple survey of the total metabolism of the Earth's crust.

After the works of V. M. Goldschmidt[1], P. Eskola[2] and others it has been a matter of course that the paragenetic transitions during metamorphism of a rock are led by chemical forces. But such transitions do not demand any considerable transport of matter. In most cases the minerals are in direct contact with each other. But as soon as the distances over which the substances must have been transported have reached a considerable amount, a mechanical transport of exceptionally mobile compounds has been suggested. For instance: injection of magmas along planes of schistosity, flow of aqueous solutions through smaller or greater fissures in the rocks, transport of gaseous compounds such as: $FeCl_8$, SnF_4 etc., and transport of the pore liquid in the so-called intergranular film. But the forces which drive this wandering of compounds have never been fully understood.

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[1] Die Kontaktmetamorphose etc. Skr. Vid.-selsk. Kristiania, Mat. Nat. kl. No. 1. 1911.

[2] Om Sambandet etc. Bull. Comm. Géol. Finland. No. 44, 1914.

The new point of view in this theory is that even the *transport of materials over long distances are governed by chemical forces*; and accordingly that we can apply the thermodynamic laws of the variation of these chemical forces in minerals (vapor tension, activity) with mechanical pressure (P), with temperature (T), with composition (X), size of minerals (Y), and with surrounding phases (Z), during the study of the metabolism of the earth's crust.

An example: the origin of a granite pegmatite several kilometres away from the parent granite is caused by the difference between the chemical activities of the granite minerals in the granite itself, and in the fissure in the country rock in which the pegmatite develops. The high mechanical pressure in the parent granite, and the low mechanical pressure in the fissure (a "mechanical vacuum") create a corresponding difference in the activities of the minerals at the several places.

Every condensed phase: the crystalline minerals, the molten magmas, and the aqueous solutions (for instance the pore liquid) consists, at any one given time, of two essentially different categories of atoms or ions: viz.: (1) Atoms possessing a so small kinetic energy that they become fixed to their environments, unable to jump out of the field of attraction of the nearest neighbours; and (2) atoms possessing sufficient kinetic energy to jump out of the field of attraction of their closest neighbours.

The former category we may name the *consolidated atoms* or the *consolidated phase*. The consolidated phase is in itself chemically inactive; only indirectly *via* the dispersed phase (see below) it can react chemically. To mechanical forces, however, it yields *en bloc* and may move, together with its contents of dispersed atoms, in the direction of the gradient of mechanical force. Examples: flow of liquids, translations in the lattice of crystals, crushing of minerals etc.

The latter category we may name the *dispersed atoms* or the *dispersed phase*. These atoms are the chemically active ones; they are able to migrate inside and outside the lattice in the direction of the gradients of the chemical activity. The dispersed atoms therefore can migrate away from one compound into another and react there, and they can exchange places with foreign atoms.

The principal forces leading the migration of the dispersed phase are differences in the chemical activities at different places in space. (The chemical activity of an element in the dispersed phase is approximately equal to the partial pressure of this element; the total chemical activity of a compound is equal to the sum of the partial pressures of the dispersed elements with which the compound is in equilibrium.[1]) But during the mechanical transport of the consolidated phase *en bloc*, the dispersed atoms are usually also moving with as a whole; this is a mechanical transport of the dispersed phase. The supposed flow of a pore liquid through the interstices between the minerals in a rock is a good example on such a mechanical transport of dispersed atoms.

Indirectly, the dispersed atoms are greatly influenced by mechanical forces; the chemical activity (the pressure of the dispersed phase, or the vapor tension) will always increase with increasing mechanical pressure acting on the condensate. This is a fact of the greatest importance for the study of metamorphism and metasomatism. It is the old Riecke's principle[2] in a new form which we will apply in the investigation of the thermodynamics of the Earth's crust.[3]

There is obviously an equilibrium between the consolidated atoms and the dispersed atoms so that dispersed atoms will consolidate if they are led to the condensate, and consolidated atoms will disperse if the dispersed atoms migrate away from the condensate.

The present theory is built up on the laws governing the variations of the mentioned equilibrium with mechanical pressure, with temperature, and with composition and size of the condensates (minerals).

We can combine the laws giving the variation of the partial pressure of dispersion (vapor tension, activity) of minerals, with mechanical pressure and temperature respectively[4]:

$$D = K \cdot e^{\frac{m \cdot (P - D_T) - \Delta F \cdot S}{RTS}} \tag{1}$$

where D is the pressure of dispersion — or the chemical activity — K is a constant if the composition of the mineral is constant; in mixed crystals, K varies continuously with the composition. ΔF is the difference in free energy between the consolidated atoms and the dispersed atoms, S is the specific gravity of the mineral, m is the atomic weight of the considered element, P is the mechanical or the total pressure acting on the mineral, T is the temperature, R is the gas constant, and D_T is the pressure of dispersion at the temperature T when the mineral is exposed to a total pressure equal to its own vapor tension.

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[1] Taylor: Treatise on Physical Chemistry, Vol. I, Edit. 2. p. 423. 1931.

[2] F. Riecke, Nachrichten Königl. Ges. Göttingen, Math.-Phys. Kl. p. 278, 1894.

[3] See also H. Ramberg, Vid.-Akad. Avhand. I. No. 3, 1944.

[4] H. Ramberg, loc. cit.

We shall not discuss this equation here; we shall only keep in mind that D usually increases with T if P is constant, and increases with P at constant T. The lower the specific gravity of the mineral, and the higher the atomic weight of the considered element, the more rapid the increase of D with pressure P.

Within confined parts of a rock a state of mutual thermodynamic equilibrium is approached during the metamorphism.

Ideally, all the solid rocks, the magmas, and the aqueous solutions in the crust should be in chemical interreaction *via* the dispersed phase which is able to migrate through the rocks in all directions, each element individually obeying the command of the chemical activities.

These ideal conditions are only developed when the system has a sufficiently long time at its disposal, and when and where the rate of chemical transport of matter in the dispersed phase is greater than the rate of mechanical transport by the consolidated phase. The mechanically flowing magmas and solutions, in which the resistance against the mechanical movement — the viscosity — is low, have therefore never reached equilibrium with the solid neigbouring rocks. But after a certain time stationary magmas and solutions must be in equilibrium with the solid surroundings *via* the dispersed phase (contact metamorphism and metasomatism).

The above mentioned ideal conditions are most completely developed in the solid rocks in the deepest parts of the crust. Because of the increase of chemical activity with pressure (and with temperature), the rate of the chemical reactions is greater than the rate of mechanical movement in sufficiently great depths. At higher levels, however, the rate of the chemical reactions decreases; in the clastosphere therefore the mechanical movements even of the rigid solid rocks will play a more important rôle than the chemical migrating processes.

At the deepest levels of the crust the volumes within which all the condensed phases (the solid rocks, the melts and solutions) are in a mutual chemical interreaction *via* the dispersed phase, are of regional dimensions (perhaps the total crust at these levels is able to interreact chemically). The volumes within which the thermodynamical equilibrium is attained, gradually narrow as successively higher levels are reached. Only the mineral units lying within a distance of a few cm apart, are in chemical equilibrium; at the top of the crust, however, even the minerals lying in direct contact with each other may represent an unstable assemblage.

Due to differences in mechanical pressure, P, temperature, T, chemical composition, X, and sizes of mineral units, Y, at different places in the earth's crust, the chemical activities — or the partial pressure of the various elements in the dispersed phase — must also vary. The locality at which the chemical activities are lowest must therefore imbibe from the dispersed phase the atoms necessary to establish the equilibrium. And these atoms must be drawn from places where the activity of the same atoms is higher. *Dispersion of minerals at places where the activities are great, Migration of the dispersed atoms towards places where the activities are less, and Consolidation there constitute the fundamental causes for the metamorphic and metasomatic processes in the Earth's crust.*

In that way the Earth's crust may be regarded as a complex precipitate in equilibrium with its solution; the precipitate corresponds to the rocks and minerals, the "solution" to the dispersed phase which is made up of those mineral-atoms that have sufficiently high kinetic energy to be able to migrate in the directions of the activity gradients.

The difference in the chemical activities at different places in the Earth can only be caused by the five variables: the mechanical pressure (P), the temperature (T), the chemical composition (X), and the size of the mineral units (Y), and the surrounding phases (Z).

The Influence of Primary Differences in Pressure.

The variation of the chemical activity with pressure is given by the equation:

$$D_P = D_T \cdot e^{\frac{m \cdot P}{RTS}} \quad \text{(see p. 100)} \qquad (2)$$

The fact that minerals have different specific gravities in combination with the effect of mechanical pressure on the activity causes an undifferentiated solid crust to try to split up by dispersion, migration and consolidation and mechanical plastic flow into monomineralic concentric layers, the heaviest below, the lighter above.[1]

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[1] H. Ramberg, loc. cit.

Even at low temperatures high mechanical pressures are able to transform solid matter into a super-dispersed phase which has a behaviour analogous to that of a highly compressed gas. In this super-dispersed state of matter all substances seem to be able to form homogeneous mixtures; i. e. when different minerals or rocks are sufficiently compressed, they will transform into *one* single homogeneous phase — the super-dispersed phase.

According to my view, thus is the state of matter encountered in the Earth's core. (This hypothesis will be discussed in a later paper.)

In an ideally differentiated Earth the partial as well as the total pressures of dispersion are at a minimum — the free energy, F, is a minimum. At any one definite level in the Earth the composition of the consolidated phase as well as the composition of the dispersed phase are constant all over the level. The composition of the consolidates (the minerals), however, varies discontinuously in the vertical direction since the minerals are arranged in concentric homogeneous layers. The composition as well as the concentration of the dispersed phase exhibit a gradual variation in the vertical direction. The concentration of all the dispersed elements increases exponentially with the depth, but the concentration of the heaviest atoms increases more rapidly than does the concentration of the light atoms.

If and when there exists a gradient in the mechanical pressure along a surface of constant temperature in the crust, then there must exist an activity gradient in the same direction. This activity gradient will cause the dispersed atoms to migrate along the gradient: the *minerals must disperse at high mechanical pressure and consolidate at low mechanical pressure.*

This is a general principle.

Now, the activities of different minerals are different under the same P, T-conditions. The elements of those minerals which have the greatest activities therefore are most concentrated — or have the greatest partial pressure — in the dispersed phase. Further, the variation with mechanical pressure of the partial activities of minerals depend on the specific gravity of the minerals and on the atomic weight of the elements. The forces which drive the dispersed atoms along a given gradient of mechanical pressure are therefore different for the various elements, and the resistance against the migration of different atoms through the rocks varies. *The different rocks in the crust must* *be regarded as semi-permeable "membranes" which to a various degree are permeable for the various atoms or ions*

In that way the elements that migrate along a mechanical pressure gradient are selected, and consequently the minerals which consolidate at low mechanical pressure are also selected.

This process I propose to call *chemical squeezing.*

In an open fissure in a solid rock at a definite level the mechanical pressure is approximately zero, in the surrounding rocks the pressure equals the weight of the superincumbent load. *Into such a mechanical "vacuum" the selected atoms in the dispersed phase must migrate and consolidate to special minerals.*

In that way I explain the origin of quartz veins, of mineral veins, of alpine veins, and of the granite pegmatites which occur within the migmatite frontier.

In the crust under an ocean depression or under a geosyncline, the pressure is less than the pressure in the same level under the surrounding continents. Consequently there exist chemical activity gradients in the directions towards such a depression in the Earth's surface; the materials from the surroundings must disperse, migrate and consolidate again at the low pressure under the depression. This process in connection with plastic mechanical flow of the solid sial, is, in my opinion, the cause of the granitisation and of the orogenesis.

Thus great bodies of granite and granodiorites are formed at low temperature in the solid state. Perhaps the problem of the origin of anorthosites can be solved in an analogous way.

During the consolidation of minerals at low mechanical pressures, the heat of dispersion (ΔF), is liberated: Around the pegmatites and the petroblastically (see definition p. 109) formed granites the liberated heat will cause an increase in the temperature of the neigbourhood, and accordingly a contact metamorhism.

The regional metamorphism can be explained by the liberation of heat during the consolidation of the granitic dispersed phase in connection with the regional granitisation.

In the same way local magma masses will form when the increasing temperature reaches the melting interval of the rocks. The maximum temperature which can occur during the consolidation of dispersed atoms is given by the difference between the mechanical pressure in the surroundings and in the place of consolidation; the greater the pressure difference, the greater the increase in temperature.

The differences between the mechanical pressures in an open fissure and in the surrounding solid rocks increases with the depth in the Earth. Perhaps this pressure difference is insufficient to cause a refusion at higher levels than at some depths in the basic substratum; this will be an interesting explanation of the great excess of basalts in relation to acid lavas.

During orogeneses however, great heat quantities are liberated during the granitisation, and refusion of acid magmas is possible.

I have explained the granitisation phenomena during an orogenesis as a migration of granitic ions from the surroundings, where the activities of the minerals are high, into the geosynclinal depression where the activities are low.

But, as a consequence of the increase of activities with the pressure, and of the low specific gravities of the granitic minerals, there ought to take place a regional granitisation all over the crust provided that the interior of the earth is not yet fully differentiated. In this case the activities of the light granitic minerals: quartz, potash feldspar and albitic plagioclase, are greater at low levels than those corresponding to the level concerned. The outer shell must therefore imbibe the ions of the granite, the ions gradually migrate outward through the crust until all the consolidated granitic materials in the Earth's interior have dispersed.

This upward migration is caused by the low density of the solid granitic material, not by the low density of the dispersed phase.

Ideally, all the light granitic solid (condensed) material of the Earth will be assembled in an outer layer; unsaturated dispersed granitic atoms (K, Si, Al, Na), however, will exist in great concentration in the rock layers farther down.

The Influence of Primary Differences in Temperature.

The relation between temperature and activity of a mineral is:

$$D = K \cdot e^{\frac{-\Delta F}{RT}} \tag{3}$$

where K is a constant, D is the partial activity of an element in the mineral, and ΔF is the difference in free energy between the dispersed atoms and the consolidated atoms of the considered element.

Accordingly there must exist activity gradients along a temperature gradient in the crust. But such an activity gradient caused only by a temperature gradient does not necessarily cause a migration of dispersed atoms through the rocks.

The migration of dispersed atoms along an activity gradient caused by a difference in mechanical pressures in solid rocks, takes place because of the very high viscosity of solids so that P-differences cannot rapidly enough be neutralized by mechanical flow.

Analogously, whether or not a migration of dispersed matter shall take place along an activity gradient caused by a temperature difference, is determined by the relation between the rate of heat conduction, and the rate of migration of dispersed atoms through rocks.

The most important primary *T*-differences in the crust are perhaps that occurring around a hot magma which is injected into cold solid rocks, and that represented by the vertical temperature gradient of the crust. The contact metasomatism is perhaps therefore partly caused by the difference in the activities of the hot magma and of the cold minerals in the surroundings. (The most important processes of contact metasomatism are probably caused by the difference in composition of the surrounding rocks and of the gases and solutions from the magma masses; the active solutions are mechanically transported into the wall rocks where they react chemically with the solid minerals.)

The vertical temperature gradient of the crust creates activity differences which may cause the dispersed atoms to migrate to the surface and solidify at higher levels. (In the undifferentiated crust the variation of activity with pressure will cause a migration of the dispersed atoms of the light minerals towards the top, and a downward migration of the heavy mineral atoms. At the same time the *T*-gradient will *accelerate* the upward diffusion of granitic substance, and *retard* the downward diffusion of basic matter. The net result being that diffusion moves more substance upward than downward. Naturally this must be compensated for by a mechanical downward flow of basic matter.)

As the dispersion, migration and consolidation, caused by differences in mechanical pressure at different places, result in an increase of the temperature where the pressure is low (p. 104), thus differences in the temperatures may, because of the consolidation of dispersed atoms at low temperature, result in a high mechanical pressure where the temperature is low (analogous to the osmotic phenomena).

The Influence of Primary Differences in Chemical Composition.

The third variable which can cause variations of the chemical activities, is the chemical composition of the system.

The activity of minerals unable to form isomorphous solutions, is constant at constant T and P (K in equations 1, 2, and 3 is a constant). The partial activity of isomorphous elements in mixed crystals increases gradually with the concentration of the considered element in the crystal.

We shall treat systems where T and P are homogeneous, i. e. do not vary within the system, but where the chemical composition is heterogeneous, i. e. the chemical composition varies from place to place within the system.

Differences in chemical activities, and consequently migrations and reactions of the dispersed atoms along the activity gradients will occur only if the phases (rocks and minerals) are paragenetically incompatible. Such a paragenetical incompatibility can be caused in different ways, viz.: (1) Mechanical transport of substances into the system, and (2) increasing or decreasing of P and/or T all over the system.

As a result of mechanical transport of condensed phases (rocks, magmas, solutions, and sediments) under conditions at which the rate of chemical reactions is low in relation to the rate of mechanical movements, associations of materials may occur in which activity differences exist. As time goes on, however, and especially if the pressure and the temperature increase, the dispersed atoms will migrate towards the places of lowest activity, and new minerals or rocks will appear.

The liberation and absorption of heat of reaction at the different places in the system (crust) will result in more or less local T-variations. Analogous to osmotic phenomena the above mentioned processes will cause minerals as well as rocks to grow against high mechanical pressure (the force of crystallization, crystaloblasthesis, petroblasthesis). (The free energy liberated during dispersion of matter at high activities, migration towards low activities, and consolidation there, can be transformed both to mechanical work and to heat.)

The most conspicuous metasomatic rock which is formed by diffusion of dispersed materials along activity gradients caused by differences in chemical composition, is perhaps the so-called reaction skarn between limestone and silicate minerals.

The contact metasomatism and the injection metamorphism are mostly caused by activity differences occurring when solutions are mechanically transported into the surroundings.

The paragenetic transitions which take place in a single rock exposed to varying *P* and *T* are exactly of the above mentioned type of reaction. *P* and *T* are homogeneous all over the system (rock); dispersion, migration, and consolidation take place because the chemical composition has become unstable.

Mixed crystals lying at the same level in the crust will theoretically react by dispersion, migration, and consolidation until they have attained a homogeneous composition (problem of the origin of anorthosites). That component of the mixed crystals which has the greatest rate of migration (the highest chemical activity) will migrate towards the components that have lower rate of diffusion (lower chemical activities). In that way the mixed crystal will not be distributed all over the level in dilute concentration in all the rocks, but it may be assembled in greater homogeneous masses.

Minerals which are paragenetically unstable, are even unstable when lying at different places in the same level in the crust. There exist activity gradients which may cause the dispersed atoms to migrate between the paragenetically unstable minerals, to react and to form new minerals (rocks).

In other words: All minerals all over the crust must, if chemical equilibrium exists, be paragenetically compatible. This conclusion is one of the weakest points of my theory. But we must remember: (1) Chemical equilibrium is not generally attained in the upper parts of the crust. (2) Only the minerals with the highest chemical activities are able to migrate over longer distances.

It is therefore reasonable that the minerals having the highest activities are mutually chemically compatible, and also that they are able to exist together with the minerals with the lowest activities without reaction. The minerals in the crust which have the lowest activities are not able to react with each other *via* the dispersed phase over longer distances; therefore it is probable that those minerals are not mutually compatible.

According to my opinion the granitic minerals: potash feldspar, quartz, albite, and perhaps some minerals rich in Fe in relation to Mg, have the greatest chemical activities (under the *P*, *T*-conditions existing in the geologically known levels of the crust). These minerals are those which are able to migrate long ways.

The basic minerals: pyroxene, hornblende, anorthite, epidote etc. seem to be less active chemically. They are usually not able to migrate over longer distances. Exposed to mechanical forces they often react by being crushed; the granitic minerals, on the contrary, usually react by dispersion and migration.

The Influence of the Size of Mineral Units.

The activity of a mineral increases with decreasing size of the mineral units in an analogous manner as does the vapor pressure of small drops.

Large crystals will therefore grow against great mechanical pressure if lying together with smaller grains of the same mineral. There exist relations between the differences in size, and the mechanical pressure against which the greatest crystal can grow (force of crystallization).

The Origin of Petroblastic Rocks.

According to the view maintained in this paper apparently "magmatic" rocks (rocks with hypidiomorphic or granitic textures) can be formed far below their melting intervals. They appear by consolidation of the dispersed phase (the atoms possessing so great kinetic energy that they are able to migrate inside and outside the lattices) at places where the activities of the minerals of the rocks are lowest.

The general requirement for the origination of rocks of "magmatic" aspect, is not that they are formed by crystallization of a liquid or a melt, but only that they are formed within the stable *P*, *T*-field of their minerals.

At the proposal of professor Tom. F. W. Barth, I will name the rocks which are formed in sub-solidus state by consolidation of the dispersed atoms without reaction with older minerals, *petroblastic rocks*. (Crystaloblasts are minerals formed in the same way.) The most typical example of petroblastic rocks is a granite pegmatite which, according to my opinion, is formed by consolidation of the dispersed atoms in a fissure (or at a place of exceptionally low mechanical pressure) in the crust.

If the dispersed phase reacts with old minerals at the place of consolidation, the newly formed rock is to be named metasomatic. (Examples: reaction skarn, some granitized rocks in which the old minerals have reacted with the dispersed phase forming granitic minerals.)

The petroblastic rocks originate by consolidation of the dispersed atoms of the solid minerals at places of low activities caused by low mechanical pressure or low temperature (pp. 103—106). (If the dispersed atoms of a melt or an aqueous solution consolidate to solid minerals, the ordinary magmatic or hydrothermal rocks occur.

Where the activities are low due to the chemical environments, the dispersed atoms must react with the older minerals; the rocks which form are metasomatic (pp. 107—108).

In this theory I assume that the transport of matter during the metasomatic and metamorphic processes in the crust takes place without the help of the so-called volatile substances or other "carriers" which commonly are assumed to be the active agents during the transport of the mineral atoms.[1] (CO_8, H_2O, Cl, F, S etc.)

An example: If a quartz dike is formed by precipitation from an aqueous solution, very great quantities of water must have been transported through the fissure. Silica has, as we know, a low solubility in water. At 336° the concentration of SiO_2 in a saturated aqueous solution is 0.22 per cent,[2] i. e. 1220 kg water must be transported through the fissure to form 1 kg quartz assuming that all the silica is precipitated from the solution. But only a smaller part of the solute (quartz) will precipitate, and then much greater quantities of water are necessary.

Further we must assume that the temperature during the whole time of precipitation must have been lower in the quartz dike than in the surroundings. The cause of the crystallization of SiO_2 from the solution must in that case be due to a cooling of the solution.

If, on the other hand, the quartz vein is formed in accordance with this theory, only those atoms which are consolidated in the quartz crystals need have been transported to the place of consolidation.

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[1] Tyrrel, G. W. The Principles of Petrology, London p. 252 and 324, 1938; and Eskola, P. in: Die Entstehung der Gesteine, Berlin p. 373, 1939.

[2] Morey, G. W. and Ingerson, E. Economic Geol. Vol. XXXII No. 5, p. 732, 1937.

And the cause of the supersaturation of the dispersed phase in the fissure is not a lower temperature in the fissure than in the surroundings, but the low mechanical pressure in the fissure in relation to the adjacent rocks. (The activity or vapor tension of every compound must always increase with the pressure; the solubility will in most cases decrease with pressure; the solubility in water of but a few compounds increase with pressure[1] if both precipitate and solution are exposed to equal pressures.) Further: the rising temperature in the quartz dike caused by the liberation of heat of dispersion will not hinder the consolidation until the temprature has reached a certain definite value higher than that in the neigbourhood.

We have seen how the reactions in the Earth can be classified in: (1) Mechanical reactions during which the condensed phases react more or less *en bloc*; the forces governing the movements are mechanical pressure gradients. (2) Chemical processes which take place in the dispersed phases; the forces driving both the dispersion and consolidation as well as the movements in the dispersed phase are chemical activity differences. The differences in chemical activities are caused by five factors, viz.: Differences in the mechanical pressure at a certain level in the crust, differences in the temperature in the crust, differences in the chemical composition of the phases (minerals) differences in the size of the mineral units, and finally, by different surrounding phases.[2]

During the development of the crust the slow chemical processes: dispersion, migration, and consolidation aim at the attainment of complete thermodynamic equilibrium. The rapid mechanical processes often disturb the equilibrium.

Acknowledgements.

I wish to express my warmest thanks to professor dr. Tom. F. W. Barth for friendly criticism of the manuscript and for interesting discussions. Thank are also due to dr. H. Wergeland, Oslo University, for discussions concerning the thermodynamic problems. I want to acknowledge the financial support that I have received from "Nansenfondet" during my work.

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[1] Gibson, R. E. Scient. Monthly Washington *46*, 1938.

[2] See notice p. 115.

Mineralogisk Institutt Oslo.
December 1944.

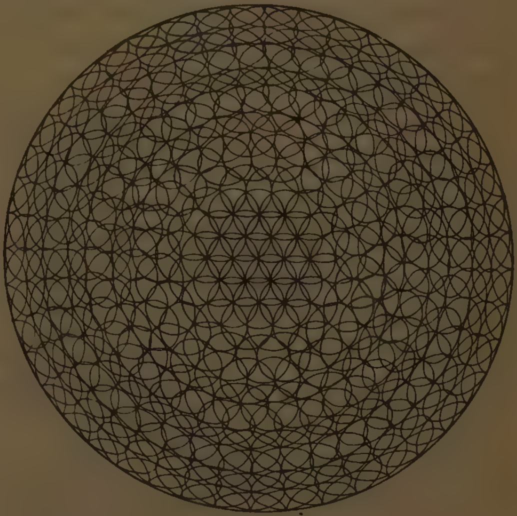
Printed, February 17th 1945.

NOTISER

A METHOD OF COUNTING OUT PETROFABRIC DIAGRAMS

The counting out petrofabric diagrams as described in the literature is carried out by means of a celluloid counter with circular holes representing 1, 1,5 or 2 per cent of the area of the circle of projection. The present writer has found it more convenient to construct a net of circles with the percentage wanted, with the distance between the centres equal to the radius of the circles. The diagram, with the projection points plotted on a tracing paper, is placed upon this net, and is covered by a second sheet of tracing paper. The numbers of points falling within each of the circles of the net can then easily be counted and are marked by corresponding figures on the upper paper. In this way the counting out of a diagram will be easier as well as more impersonal than by using a celluloid counter.

Petrofabric diagrams are plotted in an equal area projection (the Lambert projection of cartography) by use of the so called Schmidt net, and are counted out by circles as described above. This method has been criticized by Mellis[1] on the ground that the concentration in a given point of the hemisphere correctly should be measured by the number of plots falling within a spherical circle with its centre in the point. The plane circles generally used represent spherical circles only in the centre of the projection, in the peripheral parts they represent (somewhat deformed)



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[1] Otto Mellis: Gefügediagramme in stereographischer Projektion. Min.-petr. Mitt, *53*, p. 330, 1942.

ellipses on the hemisphere with the longer axes in a meridional direction. Due to this circumstance some distortion of the diagram will take place. To avoid this, Mellis proposes to represent petrofabric diagrams in the stereographic projection, in which a circle on the sphere is represented by circles in the plane projection. In the stereographic projection a diagram can correctly be counted out by circles, the radii of which, with a given percentage, increase with the distance from the centre of the projection.

There are two chief objections to the use of the stereographic projection for petrofabric diagrams. Firstly, the equal area projection has already been used for a great number of published diagrams, and secondly the stereographic projection has the disadvantage of representing equal angular distances on the sphere by greatly varying linear distances in the projection. The paper of Mellis has therefore occasioned the author to construct the net here reproduced for the counting out of petrofabric diagrams in the equal area projection, by the use of spherical circles with radius 8°,1, representing 1 per cent of the hemisphere.[1] As is seen from the figure the sphercial circles are represented by (somewhat deformed) ellipses, the difference between the axes of which increases with the distance from the centre of the projection. Within a distance of 25° from the centre the difference between the axes is less than one milimetre on a net with a radius of 10 cm, and here circles may be used without any appreciable error. Similar nets might be constructed with 1,5 per cent circles, and with 2 per cent circles, with radii of 10°,0 and 11°,5, respectively, and with circles with a radius of 5° for the constructive counting out of diagrams proposed by Mellis (*l. c.*).

By devoting some time and care every one may construct similar nets for his own use, to get fine results, however, exactness in construction as well as skill in drawing is needed. If the method here suggested should meet with general approval, it would be desirable that some firm or institution undertake the construction of the nets, which might then be reproduced and distributed for sale. The present writer is well aware, that the orientation of minerals in rocks is not a finely adjusted process so that the errors resulting from the commonly used methods are far from being serious. Yet, the use of correct methods will give satisfaction to the workers in this field, and uniformity in methods will be of importance to this branch of science as a whole.

Trygve Strand.

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[1] The area of a spherical calotte (a spherical circle with radius φ) is given by $2\pi R^2 (1 - \cos\varphi)$, where R is the radius of the sphere. $1 - \cos\varphi$ is thus the area of the spherical circle in proportion to the area of the hemisphere.

UNIT CELL AND SPACE GROUP OF CHALCOCITE, Cu_2S

In 1930, when doing X-ray work in the Mineralogical Laboratory in Cambridge, I took a number of oscillation photographs of chalcocite from Cornwall. Axes of rotation were the 3 crystallographic axes. From the photographs the following dimensions of the unit cell were derived:

$$a_0 = 11.8\ \text{Å} \qquad b_0 = 27.0\ \text{Å} \qquad c_0 = 13.4\ \text{Å}.$$

These values are not very exact; they may be uncertain by about 0.1 Å. They are in close agreement with a set of values which has been published in the meantime.[1] Because of the very large number of atoms which must be contained in this cell I did not think it possible to determine the structure from the photographs; in any case such a determination would be an extremely laborious task; and so I did not publish any account on my work chalcocite. Reviewing my old photographs I find that a fairly reliable determination of the space group is possible. The interpretation of the photographs was done by the graphical method worked out by J. D. Bernal.[2] Because of the large size of the unit cell a unique interpretation was possible only for part of the spots in each photograph. Even so I obtained a list of about 200 different sets of indices which may be regarded as well established, many of them being represented by several spots. In all of these observed sets of indices *hkl* the sum $k+l$ is even, that is: k and l are either both even or both odd; in other words: *hkl is halved if* $k+l$ *is odd.* This seems to be the only kind of abnormal spacings. The only space group in accordance herewith is C_{2v}^{14} — *Amm.* Interchanging the axes a and c also C_{2v}^{11} — *Cmm.* Q^6 — *C222* and Q_h^{19} — *Cmmm* are possible. As no faces indicating a lower symmetry than the orthorhombic holohedral have been observed on crystals of chalcocite, all these space groups but the last one are improbable. *The space group of chalcocite is therefore probably* Q_h^{19} — *Cmmm.* To be in accordance with the space group notation here adopted we have, as mentioned, to interchange the axes a and c in the usual orientation of the chalcocite crystals.

As to the number of atoms in the unit cell neither the cell dimensions nor the specific gravity of chalcocite are known with sufficient accuracy to give a unique result. Using the above cell dimensions and the specific gravity 5.8 we obtain about 93 Cu_2S in the unit cell. It is, however, highly probable that the correct number is *96*. This is supported by several facts. 96 Cu_2S in the unit cell gives a calculated density of 5.97, which is the value found for cubic Cu_2S by Tom. Barth.[3] In other words: the cubic cell, which contains 4 Cu_2S, is very exactly 1/24 the volume of the orthorhombic cell. It might be assumed that the orthorhombic cell would be built up of 24 more or less distorted cubic cells, but this does not seem to be the case. It is true that the intensities of the spots in the oscillation photographs indicate a fairly distinct pseudo cell which is 1/24 the volume of the orthorhombic unit cell, but its shape is very far from cubic. This pseudo cell is indicated by the relatively very high intensities of spots with indices like 084, 086, 344, 346, 3 12 0, 3 12 2, 3 12 4. These indices are of the form 3*n*, 4*n*, 2*n*, and the pseudo cell thus has the dimensions $\frac{1}{3}a_0$, $\frac{1}{4}b_0$, $\frac{1}{2}c_0$, or 3.94 Å, 6.75 Å, 6.70 Å. The pseudo cell is therefore pseudo-tetragonal, and also pseudo-hexagonal. (In accordance with what has been said above the product 3.94 Å × 6.75 Å × 6.70 Å is very nearly equal to $(5.59 \text{ Å})^3$, which is the volume of the unit cell of cubic Cu_2S.) — The general positions of equivalent points in the space group Q_h^{19} — *Cmmm* are sixteen-fold. This is an additional indication that the real number of "molecules" Cu_2S in the unit cell is 96 and not a number in the neighbourhood of 96, for 192 and 96 positions are readily built up of sets of 16. 12 + 6 sets of 16 will be needed, and the number of parameters to be fixed to determine the structure is probably (12 + 6) × 3, or 54.

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[1] P. Rahlfs, Z. physik. Chem. (B) *31*, 157, 1936.

[2] J. D. Bernal, On the interpretation of X-ray, single crystal, rotation photographs. Proc. Roy Soc. (A) *113*, 117, 1927.

[3] Tom. Barth, Die reguläre Kristallart von Kupferglanz. Centralblatt f. Min. etc., A, 1926, p. 285.

Oslo, Mineralogisk-geologisk museum, November 1944.

Ivar Oftedal.

NOTE ON THE METAMORPHIC DIFFERENTIATION OF SOLID ROCKS

This note should be considered as a supplement to my article in this journal *24* p. 98—111.

The activity of the given mineral usually varies with the kind and composition of the surrounding minerals (the variation of the activity with *Z*) even if the composition and the size of the given mineral and the *P*, *T* are constant. This is because of the influence of the surrounding phases on the surface tension of the considered phase. Thus, in a homogeneous rock consisting of chemically compatible minerals it may happen that the activities of the assemblage decrease if special minerals cluster concretionary together. During the metamorphism there will therefore exist activity gradients in a homogeneous rock that try to differentiate the rock by dispersion of some minerals at some places, migration of the dispersed elements towards places of lower activities and consolidation there. The variation of the activity of a mineral with the surrounding phases is thus — along with the variation of the activity with the pressure, *P*, and the size of mineral, *Y* — the most important factor to consider when treating the metamorphic differentiation theoretically. In this way I explain the occurrence of concretions of many mineral assemblages as for instance calcite in slate, chert in chalk, quartz and small pegmatite veins in gneisses, epidote, quartz and calcite in low grade amphibolites and so on. During the growth of the concretion which commonly consists of minerals containing the element with the greatest power of diffusion in the rock, the other minerals must be enriched in a zone along the boundary of the concretion.

Mineralogisk Institutt, Oslo, February 1935.

Hans Ramberg.

NORSK GEOLOGISK FORENING

Generalforsamling og møte 280. Torsdag 10. februar 1944.

Blev holdt i Vinterlandbruksskolen, St. Olavs gate 35.

Generalforsamling.

Til stede 18 medlemmer.

ÅRSMELDING FOR 1943

Siden forrige generalforsamling er 4 medlemmer avgått ved døden:

Olav Melkild. Død 23. februar 1943.
Sara Marie Villars-Dahl. Død 2. april 1943.
Jens Holmboe. Død 25. juli 1943.
Hans Glømme. Død 6. september 1943.

I samme tidsrom er innvalt 14 nye medlemmer:

251. Overgartnar Søren Steinsvoll. Universitetets botaniske hage, Trondheimsvegen 23, Oslo. $^{18}/_{2}$ 1943.
252. Mag. scient. Tore Sund. Handelshøgskolen, Bergen, $^{18}/_{2}$ 1943.
253. Stud. real. Jeanette Dahll. Valberg, Kragerø. $^{18}/_{2}$ 1943.
254. Mag. scient. Lars Lund. Geologisk museum, Trondheimsvegen 23, Oslo. $^{18}/_{2}$ 1943.
255. Lektor Arne Grønlie. Orkanger. $^{18}/_{3}$ 1943.
256. Landbruksstipendiat Jul Låg. Norges landbrukshøgskole, Ås. $^{18}/_{3}$ 1943.
257. Cand. mag. Johanne Hødal. Bergens museum, Bergen. $^{18}/_{3}$ 1943.
258. Mag. scient. Kåre Landmark. Bergens museum, Bergen. $^{18}/_{3}$ 1943.
259. Lærar Torbjørn Garthus. Garthus, Valdres. $^{27}/_{5}$ 1943.
260. Bergingeniør Carl J. G. Steenstrup. Langlia 23, Ullevål hageby. $^{27}/_{5}$ 1943.
261. Dr. Sture Landergren. Riksmuseet, Stockholm 50. $^{11}/_{11}$ 1943.
262. Cand. real. Egil Sæther. Stabekk, Bærum. $^{11}/_{11}$ 1943.
263. Konsulent Preben L. Gjertsen. Bygdøy allé 65, Oslo. $^{11}/_{11}$ 1943.
264. Bergstudent Kåre Helverschou. Villavegen 9, V. Aker. $^{11}/_{11}$ 1943.

Medlemstallet er no 154, derav 68 livsvarige og 86 årsbetalende.

Det er holdt 5 møter, 275—279, idet desembermøtet måtte innstilles, med et samlet frammøte av 147 personer. Det er dessuten holdt en ekskursjon med 18 deltakerer.

Av tidsskriftet er utkommet bd. 22 hefte 1—2.

Foreningen har gjennom O. Holtedahls innsamling fått en god økonomisk støtte for tidsskriftet. Om det blir det git en serskilt melding til generalforsamlingen.

UTDRAG AV REGNSKAP FOR 1943

Inntekt.

Beholdning, overført fra 1942:		
I kasse	kr. 131,53	
I bank	„ 430,97	
		kr. 562,50
Innkommet medlemskontingent:		
For 1940 fra 3 medl.	kr. 30,00	
„ 1941 „ 4 „	„ 40,00	
„ 1942 „ 24 „	„ 240,00	
„ 1943 „ 45 „	„ 450,00	
		„ 760,00
Renter fra livsvarige medlemmers fond		„ 186,04
Tilskot fra stat		„ 400,00
„ „ Sulitelmafondet		„ 2 600,00
Serskilte trykkingstilskot:		
Fra Nansenfondet til I. Rosenqvist: Metamorphism and metasomatism, bd. 22	kr. 550,00[1]	
Fra Det vitenskapelige forskingsfond av 1919 og Halvdan Bjørums legat til samme, ved forfatteren	„ 800,00	
		„ 1 350,00
Gaver fra private og firmaer		„ 27 100,00
Abonn. og salg av tidsskriftet		„ 5 173,50
Renter av bankinnskot for 1943		„ 54,10
		kr. 38 186,14

Utgift.

Tidsskriftet:		
Trykking Bind 21, h. 1	kr. 992,00	
— „ 21, h. 2—3	„ 3 321,50	
— „ 21, h. 4	„ 1 405,00	
		kr. 5 718,50
Klisjéer „ 22, h. 1—2	kr. 145,03	
— „ 22, h. 3—4	„ 319.22	
		„ 464,25
	Overføres	kr. 6 182,75

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[1] Til denne sum kommer de 450 kr., som Nansenfondet 1941 gav til trykking av Brit Hofseth: Levang peninsula. Se regnskap for 1941 i bd. 22, p. 213.

Overført		kr. 6 182,75
Utgifter til annonser		„ 33,30
Arbeidshjelp, porto og skrivesaker		„ 405,62
Møter og ekskursjoner		„ 30,00
Rabatt ved salg og abonnement		„ 1 235,83
Representasjon		„ 51,70
Beholdning overført til 1944:		
Kassebeholdning	kr. 259,18	
I bank	„ 29 987,76	
		„ 30 246,94
		kr. 38 186,14

Livsvarige medlemmers fond.

		Inntekt	Utgift
1943 $^{1}/_{1}$ Fondets kapital:			
Obligasjoner	kr. 3 500,00		
Bankinnskot	„ 1 415,00		
Innbet. livsv. kont. i 1943	„ 725,00		
Urørlig kapital	kr. 5 640,00		
Renter av obligasjoner 1943		kr. 126,00	
Renter av bankinnskot 1943		„ 65,04	
Forvaltning av obligasjoner			kr. 5,00
Overført renter til ordin. budgett			„ 186.04
		kr. 191,04	kr. 191,04

Regnskapet er revidert av S. Foslie og J. Helverschou og blev godkjent.

Regnskapsførerens løn blev satt til 300 kr. for 1944.

BERG- OG STEININDUSTRIENS GAVEFOND

Som det er meddelt på de tre foregående møter, april—november 1943, har O. Holtedahl foruten ved sin gave på 1000 kr. også ved en innsamling, vesentlig blandt firmaer i berg- og steinindustrien, skaffet foreningen nye midler til drift av tidsskriftet. Det er i alt kommet inn kr. 27 100,00.

Formannen gav derpå ordet til O. Holtedahl, som gav en meddelelse om innsamlingen med en liste over giverne. De er i nedenstående liste oppførte i den rekkefølge, gavene er innkomne:

Professor O. Holtedahl	kr.	1 000,00
A/S Christiania Portland Cementfabrik	„	1 000,00
A/S Bjørkaasen Gruber	„	500,00
A/S Sulitjelma Gruber	„	500,00
Folldal Verk A/S	„	300,00
Norsk Spræengstofindustri A/S	„	500,00
A/S Knaben Molybdængruber	„	500,00
Grubernes Sprængstoffabriker A/S	„	250,00
Orkla Grube-Aktiebolag	„	2 000,00
Det norske Aktieselskab for Elektrokemisk Industri	„	1 000,00
Killingdal Gruber	„	500,00
Christiania Spigerverk	„	500,00
A/S Skaland Grafitverk	„	100,00
A/S Vigsnes Kobberverk	„	100,00
Einar Stange A/S	„	500,00
Preben L. Gjertsen A/S	„	500,00
A/S Dalen Portland-Cementfabrik	„	500,00
A/S Norsk Labrador- og Granitindustri	„	250,00
Johs. Grønseth & Co.	„	250,00
A/S Granit	„	250,00
A/S Stordø Kisgruber	„	500,00
Sigurd Stave	„	1 000,00
A/S Fredriksstad Granitkompani	„	250,00
A/S Johs. Nilsen & Co.	„	200,00
Ankerske Marmorforretning A/S	„	250,00
N. S. Beer & Co.	„	250,00
A/S Wiik & Bårdsen (stillet i utsikt for flere år)	„	150,00
Ingeniør J. Helverschou	„	500,00
Norsk Diamantborings A/S	„	500,00
Lysaker Kemiske Fabrik A/S	„	500,00
Fosdalens Bergverks-Aktieselskab	„	1 000,00
Jacob Kjøde A/S	„	10 000,00
Raffineringsverket A/S, Kristiansand	„	1 000,00
	kr.	27 100,00

Gaven fra Skaland Grafitverk er innkommet ved bergmester C. C. Riiber, fra Norsk Diamantborings aktieselskap og fra Lysaker kemiske fabrik ved ingeniør J. Helverschou, som også har gitt en personlig gave, fra Jacob Kjøde, Bergen, ved cand. real. Odd Mortensen, og fra Raffineringsverket, Kristiansand, ved professor T. Barth. En særdeles verdifull bistand i arbeidet med innsamlingen er ydet av direktør Kjell Lund og ingeniør Carl J. G. Steenstrup.

Formannen refererte styrets tidligere omtalte beslutning om, av de innkomne midler, å opprette et fond på kr. 20 000,00, „Berg- og steinindustriens gavefond til støtte for Norsk geologisk tidsskrift“. Styrets forslag til statuter for fondet blev vedtatt av generalforsamlingen. Ifølge statutene skal fondet forvaltes av en komité av 3 medlemmer. Etter styrets forslag blev valt O. Holtedahl, C. Bugge og J. Helverschou for de kommende 3 år.

REUSCH-MEDALJEN

Det blev meddelt, at styret hadde tildelt KNUT FÆGRI Reusch-medaljen for hans avhandling: „Quartärgeologische Untersuchungen im westlichen Norwegen II. Zur spätquartären Geschichte Jærens“, Bergens museums årbok 1939—40, nr. 7. Fægri kunde sjølv ikke være tilstede og motta medaljen.

VALG

Styre for 1944.

Formann: K. MÜNSTER STRØM
Sekretær: P. HOLMSEN
Redaktør: I. OFTEDAL
Styremedlemmer: T. STRAND og A. HEINTZ
Varamenn: A. BUGGE og A. L. ROSENLUND.

Revisorer for 1944: S. FOSLIE og J. HELVERSCHOU.

Reusch-medaljekomité for 1944: T. BARTH og G. HOLMSEN.

Forvaltingskomité for Berg- og steinindustriens gavefond for 1944—46: O. HOLTEDAHL, C. BUGGE og J. HELVERSCHOU.

Dermed var generalforsamlingen slutt.

Statuter for Berg- og steinindustriens gavefond til støtte for Norsk geologisk tidsskrift.

§ 1. Berg- og steinindustriens gavefond til støtte for Norsk geologisk tidsskrift er stiftet 1943 på grunnlag av gaver fra en rekke personer og bedrifter, vesentlig innen berg- og steinindustrien. Fondets grunnkapital er kr. 20 000,00.

§ 2. Fondet forvaltes av en komité valgt av Norsk geologisk forening. Dets midler skal anbringes således som for offentlige stiftelsers og legaters midler bestemt. Fondets kapital må ikke røres. Den kan økes ved avsetning av den årlige renteavkastning, ved gaver eller på annen måte. Komitéen skal avlegge årsmelding og regnskap for generalforsamlingen.

§ 3. Komitéen skal bestå av 3 medlemmer som velges av generalforsamlingen for 3 år.

§ 4. Forandringer i disse statuter kan bare gjøres på samme måte som fastsatt i lov for Norsk geologisk forening § 10 for forandringer i samme lov.

Medlemsmøte 280.

Til stede 18 medlemmer og 2 gjester.

FOREDRAG

ISAK UNDÅS: *Raene i Østfold og de marine grenser ved dem.*

Vil seinere bli publisert i Norsk geogr. tidsskr. Nogen lokaliteter er allerede beskrevne i Norsk geogr. tidsskr. 9, 1942, p. 33.

I ordskiftet etter foredraget deltok G. Holmsen, H. Rosendahl, O. Holtedahl og foredragsholderen.

Møte 281. Torsdag 23. mars 1944.

Møtet kunde atter holdes i Vitenskapsakademiet, Drammensvegen 78.

Til stede 21 medlemmer.

Formannen meddelte, at foreningens medlem, konservator A. Nummedal, døde 7. mars.

H. ROSENDAHL holdt minnetalen om A. NUMMEDAL. Trykt i dette bind p. 89.

INNVALG

265. Ingeniør ASLAK KVALHEIM. Geologisk museum, Oslo.
Etter forslag av K. Kristoffersen og I. Oftedal.

FOREDRAG

HANS RAMBERG: *Fysikalsk-kjemiske reaksjoner i den faste jordskorpe.*

Trykt som avhandling i dette bind p. 98.

I ordskiftet etter foredraget deltok A. Bugge, T. Barth, S. Foslie, H. Rosendahl, J. Bugge, E. Sæther, P. Holmsen og foredragsholderen.

Møte 282. Torsdag 13. april 1944.

Til stede 21 medlemmer.

FOREDRAG

IVAR OFTEDAL: *Skandium i biotit, et geologisk termometer.*

Trykt som avhandling i bd. 23 p. 202.

I ordskiftet etter foredraget deltok S. Foslie, H. Ramberg og foredragsholderen.

TOM. F. W. BARTH: *Systematisk petrografi av Oslo-feltets eruptivbergarter.*

Blir trykt i Vit.-akad. skr.

I ordskiftet etter foredraget deltok S. Foslie, T. Strand, E. Sæther og foredragsholderen.

Møte 283. Torsdag 4. mai 1944.

Til stede 26 medlemmer, 1 gjest.

Formannen meddelte at det vil bli holdt en geologisk ekskursjon til Hadeland under ledelse av dr. Leif Størmer.

Statsgeolog dr. ARNE BUGGE holdt minnetale over foreningens avdøde medlem direktør Sverre Blekum. Trykt i dette bind p. 40.

FOREDRAG

GUNNAR HOLMSEN: *Kort meddelelse om et kunstig frembrakt leirfall.* Med lysbilleder.

JOHANNE HØDAL: *Et anorthositkompleks på Vossestrand.*

Blir trykt som avhandling i dette bind h. 3—4.

I ordskiftet etter foredraget deltok T. Barth, A. Bugge og foredragsholderen.

Ekskursjon 28. Mandag 7. juni 1944.

17 deltakerer.

Med tog til Lunner på Hadeland, hvor det under ledelse av L. Størmer blev sett på det område som han har beskrevet i Norsk geol. tidsskr. *22*, 74—91, 1942, under titel: Kaledonisk dekke-tektonikk på Hadeland. Til slutt blev det også sett på kalksedimenter som utfelles i innsjøer beskrevne av K. Münster Strøm i Oslo Vid.-Ak. Skr. 1941 no. 7 og Tidsskr. norske landbr. 49, 1942. Tilbake til Oslo med tog fra Lunner.

Møte 284. Torsdag 19. oktober 1944.

Til stede 17 medlemmer, 5 gjester.

L. STØRMER: *Minneord over Carl Wiman.*

Professor Carl Wiman, som ble medlem av vår forening i 1923, avgikk ved døden i sommer i en alder av 77 år.

Carl Johan Joseph Ernst Wiman var født i Husby Odensala i Stockholms Län 10de mars 1867. I 1895 tok han doktorgraden i Upsala på et paleontologisk arbeide, og samme år fikk han et dosentur i paleontologi ved universitetet. I 1911 ble der opprettet et personlig professorat for ham i paleontologi, og i 1922 ble dette overført til et ordinært professorat i paleontologi og historisk geologi. Wiman hadde denne stilling like til sin 70-årsdag da han gikk av som emeritus.

Wiman offentliggjorde sitt første paleontologiske arbeide i begynnelsen av 90-årene. 90-årene representerer på en måte et vendepunkt i svensk paleontologisk forskning. Sverige hadde gjennom en årrekke hatt fremrakende paleontologer. Vi kan bare nevne navn som Linnarsson og Angelin blant paleozoologene. Men deres arbeide hadde først og fremst gått ut på å gi en systematisk beskrivelse av fossilene i relasjon til deres geologiske opptreden. I begynnelsen av 90-årene gikk fossilstudiet over til å bli en rent zoologisk vitenskap hvor moderne zoologiske metoder fikk en utstrakt anvendelse. Det var i første rekke graptolitene som ble gjenstand for disse nye undersøkelsesmetoder, og det var de svenske forskere Holm, Törnquist og Wiman som her gikk i spissen.

Riktignok hadde Gümbel allerede i 1878 forsøkt å isolere graptolitene fra stenen ved hjelp av syrer, men gode resultater ble først oppnådd av Holm som klarte å etse graptolitene ut av kålken på en slik måte at han kunde studere de fineste detaljer i overflaten av skallet eller peridermen. Törnquist foretok serieslipninger av graptoliter, men det var først den unge og entusiastiske forsker Carl Wiman som fullt ut forstod å utnytte de zoologiske metoder. Graptolitene ble etset ut av kalken ved hjelp av forskjellige syrer, dernest bleket og innleiret i paraffin for deretter å bli snittet i mikrotom fullstendig som et resent zoologisk objekt.

I en serie av arbeider fra 1893—1901 offentliggjorde Wiman de viktige resultater av sine studier over graptolitenes morfologi. Hans resultater som bygger på nøyaktige iakttagelser og fantasirike, men forsiktige slutninger, er av blivende verd for forskningen. Dr. Bulman som er en av de mest kjente eksperter på graptoliter sier i sin omtale av Wiman's graptolitarbeider, at selv om Wiman bare hadde skrevet sine 2 første arbeider fra 1893 „his work would rank very high in graptolite literature“.

Samtidig med studiet av graptolitene publiserte Wiman betydelige arbeider over kambrosilur-faunaen, særlig fra Jämtland og Nordbaltiske områder. Han har også levert mindre arbeider over forskjellige hvirvelløse dyregrupper, blant annet triboliter, og i 1914 publiserte han et større arbeide over Karbon-brachiopoder fra Spitsbergen.

På den svenske Sydpolarekspedisjon i 1901—1903 fant Nordenskiöld på Seymourøya noen vertebratfossiler i lag av gammel-tertiær alder. Disse fossiler som fortrinsvis var knokkelrester av fugl, ble i 1904 overlatt Wiman til bearbeidelse. Dette førte Wiman over i studiet av fossile hvirveldyr. I 1908 kom han sammen med Bertil Högbom tilbake fra en ekspedisjon på Spitsbergen. Han sier selv at de kom tilbake med overbevisningen om at der fantes brukbart fossilmateriale av vertebrater i Devon og Trias på Spitsbergen. I flere arbeider beskrev Wiman stegocephaler eller panserpadder fra Trias i Sassenfjorden. Men særlig interesserte han seg for fiskeøglene eller ichtyosauriene som ble funnet deroppe. Dette førte til at han også beskrev utenlandsk materiale av liknende fossile øgler.

Før Wimans tid var det dårlig bevendt med fossile vertebrater i Upsala Universitets samlinger. Med en utrettelig energi begynte han nu en komplettering av samlingene. Flere mindre saurieekspedisjoner ble sendt til Spitsbergen. Med privat og offentlig støtte gikk han til innkjøp av verdifulle eksemplarer i utlandet. I Amerika hadde han den kjente „fossil hunter“ Sternberg til å samle dinosaurier for seg i de berømte krittlagene i New Mexico. Sin største tilvekst fikk institusjonen gjennom J. G. Anderson's enorme samlinger av fossile pattedyr fra Tertiær i Kina. Det store materiale som bestod av flere hundre kasser krevet en lang og inngående preparering og montering.

Det gjaldt ikke bare for Wiman å få samlet alle disse verdifulle vertebratfossiler til universitetet. Med entusiasme beskriver og tolker han sine fossiler. Et arbeide behandler hjernen og sanseorganene hos en fossil fisk (*Tremataspis*) fra silur. Flyveøglene fanger særlig hans interesse. Disse eiendommelige skapninger beskrives og diskuteres i en rekke arbeider, og vi hørte ham berette om dem i et ekstraordinært møte i Norsk Geologisk Forening her i Oslo 6te oktober 1925. Wiman viser seg også som en utmerket kjenner av dinosauriene fra kritt-tiden. Det store materiale av pattedyr fra Kina, særlig hornløse neshorn, er mest blitt bearbeidet av Wiman's elever. Selv har han i de senere år vært interessert i fossile kjempefugl som *Aepyornis* fra Madagaskar.

Som vi ser spenner Wiman's 50-årige paleontologiske forskning over et meget vidt felt. I sine tidligere år gjorde han en stor innsats i studiet av hvirvelløse dyr og faunaer, og i de senere år nådde han frem til å bli en av de beste kjennere av forskjellige vertebratgrupper.

Når vi omtaler Wiman's innsats må vi også nevne det verdige minnesmerke han har satt etter seg i form av den nye Paleontologiska Institutionen i Upsala. Gjennom mange år hadde han kraftig og stedig fremholdt for myndighetene hvor nødvendig det var å få en ny moderne bygning til bearbeidelse og montering av de store samlinger som universitetet etter hvert hadde erhvervet. Resultatet ble den nye Paleontologiska Institutionen, en flott bygning som danner en verdig ramme om de verdifulle samlinger som for en stor del er tilveiebragt gjennom Wiman's initiativ og dyktighet. For oss som har hatt anledning til å besøke Institutionen har det vært en opplevelse å bli vist rundt av Wiman og høre ham berette på sin elskverdige, men samtidig livfulle måte om de forskjellige merkelige fossilformene.

Carl Wiman's vitenskapelige innsats trer tydelig frem gjennom hans store paleontologiske produksjon, men det bør også nevnes at han gjennom sitt arbeide har vist at paleontologien kan hevde seg som en selvstendig biologisk vitenskap og samtidig beholde kontakten med den geologisk-stratigrafiske forskning.

FOREDRAG

CHR. OFTEDAHL: *Høgtemperatur-plagioklas i Oslofeltet.*

Foredragsholderen beskrev først den utvikling som plagioklasbestemmelsene har undergått i de siste to menneskealdre. Det nyeste på dette område er den oppdagelse som ble gjort av A. Köhler i Wien. Der er to serier plagioklaser, en høgtemperatur- og en lavtemperaturserie. Den første finnes i vulkanske bergarter som har størknet ved relativt høg temperatur, mens den annen serie er den vanlige som finnes i dypbergarter og metamorfe bergarter. De kurver som er utarbeidet for plagioklasbestemmelser er utelukkende laget på grunnlag av lavtemperaturfeltspat, slik at når disse kurver ble anvendt til bestemmelse av lavaplagioklaser, fikk man gale verdier eller resultatene stemte ikke. Köhler og hans medarbeidere har nu utarbeidet kurver også for den nye serie. Da foredragsholderen fikk to distinkt forskjellige resultater ved å bestemme en rombeporfyrplagioklas med Fedorow-bord og med immersjonsvesker, og lette etter en forklaring på dette i litteraturen, fant han Köhlers oppdagelse. Ved hjelp av de nye kurver kunde så plagioklasinnsprengninger i akeritporfyrer og rombeporfyrer bestemmes, idet de tilhører høgtemperaturserien. En artikkel om dette med nærmere data over målingene finnes i dette bind av Norsk Geologisk Tidsskrift.

Så fortsatte foredragsholderen med nærmere omtale av rombeporfyrenes fenokrystaller. Ifølge Brøgger består disse av „Natron-mikroklin“ (sjeldnere „Natron-orthoklas“) og i noen tilfelle av plagioklas (oligoklas). De består imidlertid alle av plagioklas og er antipertitisk bygget. Dette forhold ble funnet av E. Ljungner og er beskrevet i et arbeide over den svenske Skagerak-kysts morfologi (1927). Foredragsholderen har målt sammensetningen i alle de analyserte typer av rombeporfyrlava. Det viser seg da at innsprengningenes An-gehalt avtar med bergartenes tiltagende surhetsgrad. De basaltiske lavaer (såkalte „essexit“-lavaer) har innsprengninger med mer enn 50 An, og de nær beslektete rektangelporfyrer dekker intervallet 50—40 An. Rombefeltspaten i de basiske rombeporfyrer (Brøgger: Kjelsåsitporfyrer) ligger også omkring 40 An, de andre fordeler seg i intervallet 40—30 An, med noen få eksempler ned mot 20 An, d. v. s. ned mot larvikiten, idet dens rombefeltspat inneholder 18—25 An.

Til slutt ble Tyveholmgangen nærmere omtalt. Denne rombeporfyrgang har blitt ansett som prototypen på rombeporfyr, og den har blitt undersøkt av en rekke geologer, og dens feltspatfaser er blitt analysert gjentatte ganger. Brøgger har fått både rombefeltspaten og grunnmassefeltspaten analysert (Die Eruptivgesteine . . . , bind VII, 1933), men ved siden av analysekolonnen har han en kolonne med „den sannsynlige sammensetning av innsprengningene“. Her er Na_2O-innholdet revidert for å passe med norm-beregningen. Den annen analyse har Brøgger gjengitt nesten uforandret i kolonnen „sannsynligste sammensetning av grunnmassefeltspaten“. Analysen er imidlertid gal. Begge feltspatfaser er blitt analysert ved Mineralogisk Institutts laboratorium av frk. A. Thorkildsen, og de er derved blitt sikkert bestemt. Rombefeltspaten er en antipertitt med opp til 50 % alkalifeltspat; plagioklasfasen inneholder 30 An. Grunnmassens feltspattavler består derimot av natronorthoklas.

I ordskiftet etter foredraget deltok T. Barth, E. Sæther og foredragsholderen.

KAARE MÜNSTER STRØM: *Landskapsformene i Lofoten belyst ved flyfotografier.* Med lysbilleder.

Foredragsholderen betonte at det etter de utmerkede undersøkelser over Lofot-øygruppens geomorfologi som forelå av Helland, J. H. L. Vogt, Th. Vogt og Ahlmann ikke kunde fremkomme noe vesentlig nytt. Hovedsaken var å demonstrere den ypperlige oversikt og innsikt som fremkommer ved iakttagelse fra luften, i dette tilfelle ved fotografier opptatt av Widerøe's Flyveselskap A/S. Dessuten hadde han ved opphold 1935 på Moskenesøy, 1937 på Vestvågøy og Røst, 1939 på Austvågøy og Moskenesøy, og ved kartstudier kunnet supplere og systematisere de eldre iakttagelser.

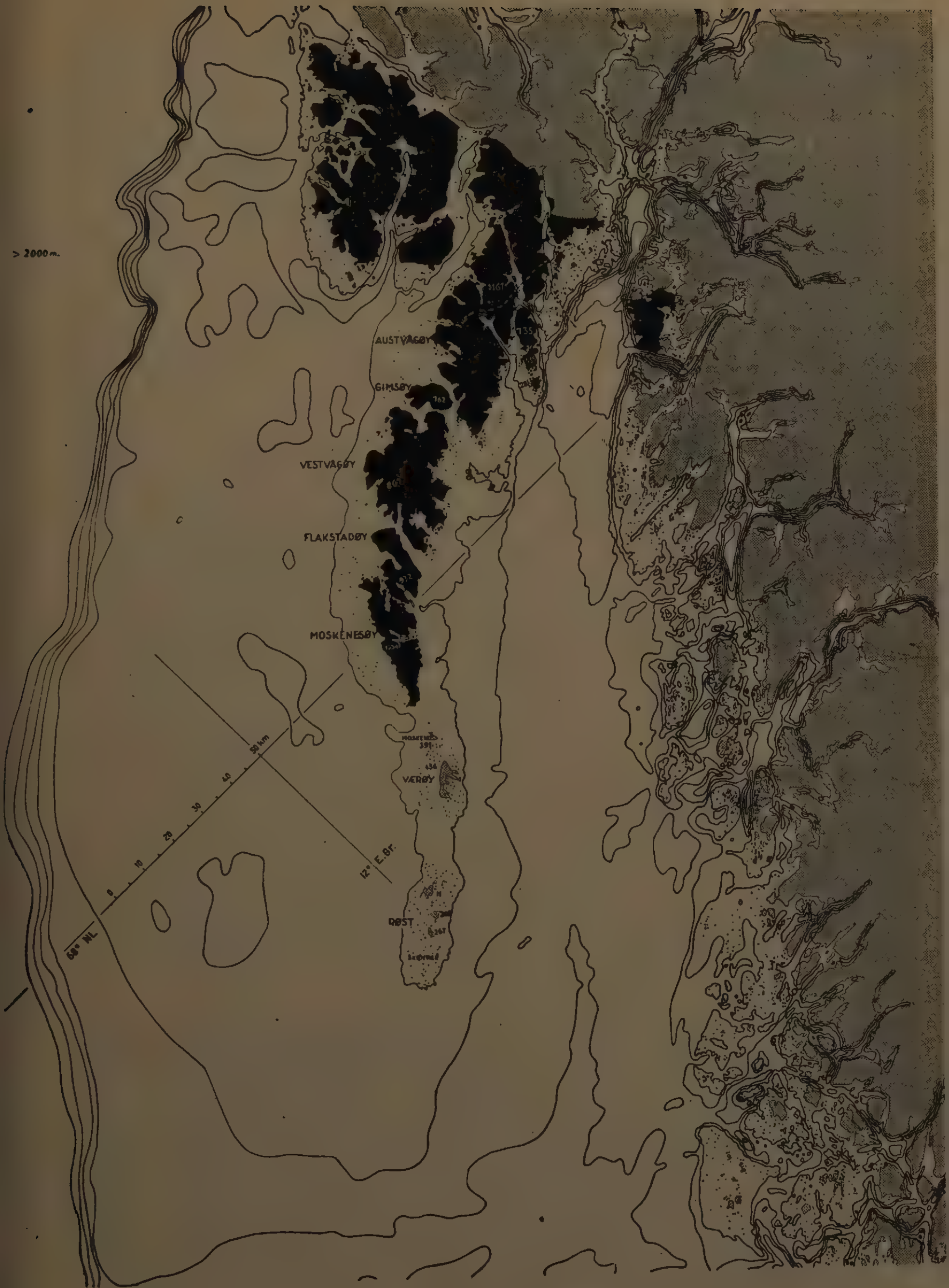
Geomorfologisk må det skilles mellom den egentlige Lofot-øygruppe og Værøy—Røst-øygruppen. En vanskelighet for den geomorfologiske analyse ligger i det å skjelne mellom innflytelsen av substrat og prosess. Substratet er i det egentlige Lofoten dyperuptiver. Th. Vogt har elegante iakttakelser over de forskjellige dyperuptivbergarters innflydelse på detaljformene i landskapet, men den generelle karakteristikk av substratet må være at det er sammensatt av harde og seige bergarter. I motsetning hertil er substratet i Værøy—Røst-øygruppen lite motstandsdyktig gneis, som i alle fall på Røst de fleste steder lett faller fra hverandre i planparallelle horisontale lag. Ved denne forskjell i substrat overdekkes i høy grad forskjellen i prosess, nemlig at Værøy—Røst i motsetning til selve Lofoten under siste istid må antas ikke å ha hatt egentlige breer.

I den egentlige Lofot-øygruppe kan man skille mellom fem forskjellige geomorfologiske elementer:

1. *Den „gamle“ landoverflate*, hvis alder ikke nøyere kan fastsettes, men som er eldre enn siste nedisning. Den har hatt en avrundet, men delvis sterkt bølgende topografi, og har ikke på noen måte vært et peneplan. Av denne overflate finnes det rester som en høyeste toppflate der hvor erosjonen ikke har ødelagt den. Det er meget usannsynlig at der noe steds finnes rester av den tertiære overflate, men med Th. Vogt antok foredragsholderen at den „gamle“ overflate for en stor del er konform med den tertiære.

2. *Den småkuperte isavrundete overflate* langs fjellfoten og oppover fjellsidene opp til større høyde i innre enn i ytre Lofoten. I innre Lofoten har den vært overflytt av innlandsisen, i ytre Lofoten av is fra lokalbreer.

3. *De konkave innsnitt i landmassen* uterodert ved lokale breer, vesentlig botnbreer, noen steder dalbreer. Botnene er gjennomgående,



Lofotøygruppen med omgivende hav. Områder som inntas av Lofoteruptivene er svarte, øvrige bergarter med raster. Dybdekurver med 100 m ekvidistanse. (Etter O. Holtedahl: Dybdekart over de norske kystfarvann *3*, 1940.) Største høyde på hver øy er angitt i m. Grensen over Austvågøy er den administrative grense for Lofoten.

og især på Moskenesøy sterkt overfordypet. Der er nu i dem fjelldemmete innsjøer med dyp opp til over 170 m. På sine steder, mest øyensynlig ved Trollfjord og Trollfjordvatn, har botnbreer langs sprekkesoner i hurtig bakoverskridende arbeide uterodert daler, som står i klar motsetning til de runde botner i mere massive bergartspartier, og til de fjorddaler som ved konfluens av botnbreer til dalbreer med sterkt øket erosjonsevne er uterodert i de massive bergartspartier. Flere av sundene mellom øyene ligger utvilsomt i lignende sprekkesoner, men her har erosjonsprosessen vært av mere blandet karakter.

Botninnsnittene er det karaktergivende geomorfologiske element. Spesielt må Moskenesøy karakteriseres som et „fretted upland", hvor det vesentlig finnes salegger og tinder mellom botnene og bare på få steder rester av den „gamle" overflate. På de innre øyer spiller de øvrige elementer en større rolle enn på Moskenesøy.

4. *Strandflaten* av utseende som ellers langs Norges vestkyst, og av usikker alder. Den haveksponerte halvdel av Moskenesøy er som påvist av Th. Vogt borterodert ved i alle fall vesentlig marine agenser, som i synkende nivåer må ha vært virksomme siden begynnelsen av den tertiære landhevning. En abrasjonsplattform som i et dyp fra 30 til 60 meter inntar det borteroderte område, må være eldre enn siste istid.

5. *En (postglasial) abrasjonsflate med tilhørende cliffs og huler* på yttersiden av Moskenesøy. Den finnes som påvist av Th. Vogt særlig utarbeidet i de minst motstandsdyktige dyperuptiver (labradorsten). Da den overskjærer en botn, må den for en vesentlig del være postglasial.

De ytre øygrupper Værøy og Røst har ikke under siste istid vært nediset, heller ikke lokalt, men toppflatene har hatt et breaktig dekke av sne og is, Værøy dessuten ubetydelige breer langs fjellfoten. De geomorfologiske elementer på disse øygrupper er: 1. Den gamle landoverflate, 4. Strandflaten, 5. Inter-, Würm- og postglasiale abrasjonsflater med cliffs, huler og (på Værøy) cirkusser. I hvilken grad de omtalte småbreer på Værøy kan ha bidratt til erosjonen av cirkussene er usikkert, men i betraktning av den lite motstandsdyktige bergart kan selv små breer ha hatt betraktelig erosiv virkning som imidlertid nu er overdekket av abrasjonen. Elementene 4 og 5 overdekker for øvrig hinannen.

Litteratur: 1897 A. Helland: Lofoten og Vesteraalen. — N. Geol. Unders. *23*; 1907 J. H. L. Vogt: Über die schräge Senkung und die spätere schräge Hebung des Landes im nördlichen Norwegen. — N. Geol. Tidsskr. *1*, 6; 1907 J. H. L. Vogt: Über die lokale Glaciation an den Lofoteninseln am Schlusse der Eiszeit. — ibid. *1*, 7; 1913 Th. Vogt: Landskapsformene i det ytterste av Lofoten. En geomorfologisk studie. — N. Geogr. Selsk. Årb. *23*. 1911—1912; 1919 H. W. Ahlmann: Geomorphological Studies in Norway. — Geogr. Ann. *1*; 1936 M. Byland: Glazial-morphologische Untersuchungen auf Lofoten und Vesteraalen. — Zürich 1936; 1938 K. M. Strøm: Moskenesøy. A Study in High Latitude Cirque Lakes. — Skr. Vid.-Akad. Oslo, 1, 1938, 1; 1940 O. T. Grønlie: On the Traces of the Ice Ages in Nordland, Troms, and the South-Western Parts of Finnmark in Northern Norway — N. Geol. Tidsskr. *20*.

Trykt 26. februar 1945.

ROCKS OF THE ANORTHOSITE KINDRED IN VOSSESTRAND, (NORWAY)

BY

JOHANNE HØDAL

With 29 figures in the text, 1 geological map.
Sammendrag på norsk.

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> *There is a pleasure in the pathless woods,*
> *There is a rapture on the lonely shore,*
> *There is society where none intrudes,*
> *By the deep sea, and music in its roar,*
> *I love not Man the less, but Nature more*
> *From these our interviews, in which I steal*
> *From all I may be, or have been before*
> *To mingle with the universe and feel*
> *What I can ne'er express, yet cannot all conceal.*
>
> Byron.

PREFACE

This work has been carried out for the Mineralogical-Geological Institute of the Bergens Museum.

The investigations in the field have been carried out during the summers 1942—1944 of which a total of 13—14 weeks were spent on field work proper. The Mineralogical-Geological Institute of the Bergens Museum has defrayed the travelling expenses.

The working-up of the collected material also has been performed at the Mineralogical-Geological Institute of the Bergens Museum, to begin with under the guidance of Professor dr. Niels-Henr. Kolderup. He joined me in the field for a few days of the summer 1942. I am also indebted to him for having obtained the opportunity to begin this work. I am glad to use this opportunity of thanking Professor Kolderup especially, also because it is first of all due to him that I became interested in geology — an interest which continued to grow under his skilful and kind guidance during the first years of my studies.

During the absence of Professor Kolderup I have worked under the direction of Amanuensis Anders Kvale who has supervised the work during most of the time. I am very much indebted to Amanuensis Kvale for his continuous interest in my work and his unfailing readiness to discuss the problems with me. He has with gracious patience answered all my questions, and placed his insight and good advice at my disposal.

During my stays in Oslo I have had the opportunity of discussing many problems with Professor Tom. F. W. Barth. I would here like to offer him my sincere thanks for the kindness he has shown me.

During the work at the Mineralogical-Geological Department of the Bergens Museum I have been able to use all the equipments of the Institute. The micro-photos have been taken by Miss Ellen Irgens, who also has given me valuable advice as to the drawing of maps and figures. My sister Mrs. Anna-Marie Bleiklie has kindly assisted me with the drawings. Preparator Martin Søgnen has made some 180 thin sections for this work.

To all these I offer my sincere thanks.

Finally I would like to thank the inhabitants of Vossestrand for all the hospitality and kindness they have shown me. That I everywhere have been so kindly received and so well boarded has contributed to making the work in the field easier to me, and more pleasant. There are many I would like to mention, but I think I do not do injustice to any one by expressing my particular thanks to farmer Torstein Skjervheim. Not only was I always kindly received in his home, but he also gave me valuable help by his detailed knowledge of local conditions and of the geology of the place.

INTRODUCTION

The area investigated is situated on the quadrangle Vossestrand C 32 V., and is indicated on the map sketch, Fig. 1. It comprises the greater part of the rural district of Vossestrand, in the north a bit of Vik is included, and farther to the west a bit of Evanger. The adjoining districts are otherwise Leikanger, Aurland, and Voss.

The area is intersected by a series of valleys, the majority following the strike direction. Farthest to the east Brekkedalen ascends from Stalheim. To the west runs Uppheimsdalen which is continued southwards by Langedalen. Farther to the west is Myrkdalsvannet in Myrkdalen which is continued by Kvassdalen. On both sides of this stand the highest mountains, Bjørndalskamben 1402 m and Finnbufjellet 1336 m. The valleys are as a rule narrow and steep. Only Myrkdalen is inhabited. Still farther west the area is not so intersected, it is an uneven mountain plain of the constant altitude of 900—1000 m, with peaks 1100—1200 m high.

The two largest lakes are Uppheimsvannet and Myrkdalsvannet of 3.9 and 1.9 km² respectively. The habitations are concentrated around them and southwards along the main road to Voss. But otherwise pastures, woodland, and outfarm land occupy the greater part of the area. The timber line is low here, generally at 600—700 m, and the greater part of the field lies above, and is bare rock.

The field is situated in the Caledonian Mountain Range on the north-west side of the Sogn-Jotunheim syncline in the most westerly corner of the large trust masses. The dip is everywhere easterly, towards the syncline, and the lowest levels are farthest to the west. The lowest layer consists of Cambro-Silurian mica schist. Trust into this are several massifs of quartzite and quartzitic schists.

Farther to the east, above the mica schist is a complex of the rocks of the anorthosite kindred. These rocks occupy the greater part of the field, and chiefly these will be described here. The borders towards the mica schist are secondary; I have shown that the rocks have been thrust into their present position. The strike direction is NW—SE, and the dip ENE. The usual Caledonian direction NE—SW is found in the quartzites and quartzitic schist situated farther west, partly in the mica schist as well. The differences

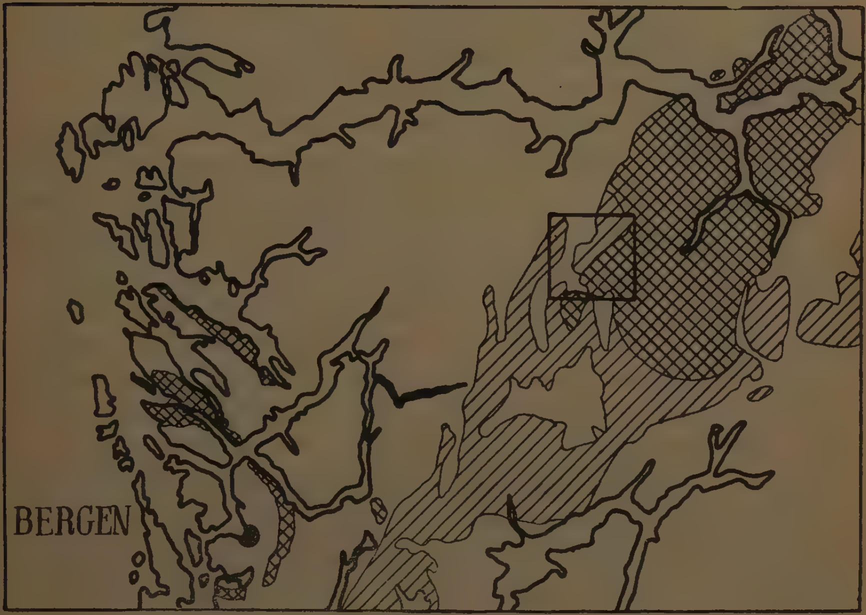


Fig. 1. Key map. The area investigated is indicated by a square. Hatched: granite, hypersthene syenite, "monzonite". Cross-hatched: anorthosite. Distribution of rocks after V. M. Goldschmidt (1916).

of strike direction are clearly reflected in the topography, the valleys following the strike direction.

Stretching, fold-axes, and thrust direction are subparallel to the NW—SE direction. The deviation between strike direction and stretching is generally 10—15°.

The greater part of the area was mapped by J. Rekstad in the summers of 1906—1908 (Rekstad 1909). C. F. Kolderup's diary from a trip from Voss to Stalheim in 1899 is available, and finally N. H. Kolderup and A. Kvale visited Myrkdalen in 1933. A sample from Ulvund was determined as mangerite.

SHORT DESCRIPTION OF THE SCHISTS

1. Mica schist is common over the whole north-western area, besides in Langedalen, in the most southerly part of Uppheimsdalen and southwards along the main road to Voss. From here a narrow streak wedges out westwards to Skreia and northwards to Kaset and Nyestøl.

Fossils have not been found here, but there is connection southwards with mica schist in Voss, and northwards with that in Sogn, so that the age certainly must be Cambro-Silurian.

Of appearance it varies somewhat; some types are rather phyllitic — it may be mentioned that Rekstad (1909) termed the rocks phyllite — but taken as a whole the metamorphism has gone so far that mica schist would be the correct term. It may contain so much graphite that it is black and glistening and blackens the fingers. These types generally have small-flaked mica. Other types contain little or no graphite; they have more coarse-flaked mica, and are often brown-coloured from iron solutions. Common to all types is that they are full of quartz lenses which are twisted and contorted. Dikes of quartz have also been observed.

Most often the metamorphism has gone so far that no traces of bedding are seen, but a thin section of a phyllitic schist at Framnes has shown parallel bands rich in graphite, which must represent a bedding. *The schistosity is here parallel to the bedding.*

The strike varies somewhat, but runs most frequently NS, and represents an interposition between the NW—SE direction of the rocks of the anorthosite kindred and the NE—SW direction of the quartzitic schists. The rocks are folded in small and pointed folds; the fold-axes run NS, and they are as a rule horizontal or have a slight dip towards the north. A total of 11 thin sections from mica schist area have been examined, showing petrographically no great variations. The chief minerals are *quartz* and *muscovite*. Quartz shows as a rule undulating extinction. Muscovite has been determined in a sample from Hielven.

The optic angle has been measured with micrometer ocular, $2E = 60°$. $n_\beta = 1.595$, ɔ: $2V = 36°$.

According to Winchell (1933) this gives the following composition:

58% Phengite $H_4K_2(Mg, Fe)Al_4Si_7O_{24}$
27% Muscovite $H_4K_2Al_6Si_6O_{24}$
15% Fe''' Muscovite $H_4K_2Fe_2'''Al_4Si_6O_{24}$

Some brown *biotite* is generally found together with the muscovite. Often it occurs together with *chlorite*, of which it is clearly a transformation. Graphite is found in most types, but may be completely missing. Calcite is found in some thin sections in considerable quantities, but may also be completely lacking. A little albite is found, particularly in the boundary regions. Ore is common, mostly pyrite, but hematite is also found. Accessories are zircon and tourmaline.

Where the altitude above the sea-level is not too great, the mica-schist possesses a rich vegetation, and as a rule it may be distinguished at a long distance from the harder rocks near by. The surface is uneven with rounded forms.

2. Quartzites and Quartzitic Schists. The great massifs thrust into the mica schist are mainly composed of quartzites and quartzitic schists. Part of these schists seem to be altered quartz porphyries. Small areas of amphibolite are also found. The boundary towards the mica schist is uneven and lobated, a typical thrust boundary. Farther south on the quadrangle Voss a succession of important slate quarries are situated on the border between the mica schist and these rocks (Kolderup 1933). The rocks are hard and resistant against weathering, and form, as a rule, high mountains. By their light hue and scanty vegetation they are sharply distinguished from the mica schist (Fig. 2). Rekstad (1909) terms these rocks "granulite and quartz rock", while Reusch on the Voss quadrangle (1905) distinguishes between "quartzite and fine-grained gneiss which often resembles quartzite in appearance". But the variations are so great that it is impossible to apply a common denomination. Among the samples I have examined by means of the microscope, quartzites and quartzitic schists predominate. But there are also types displaying under the microscope porphyric texture with larger grains of quartz and feldspar in a fine-grained ground mass. As a rule the rocks are exceedingly fine-grained, and microscopically it is very difficult to keep the various types apart. Stratigraphically they are situated between the mica schist and the rocks of the anorthosite kindred.

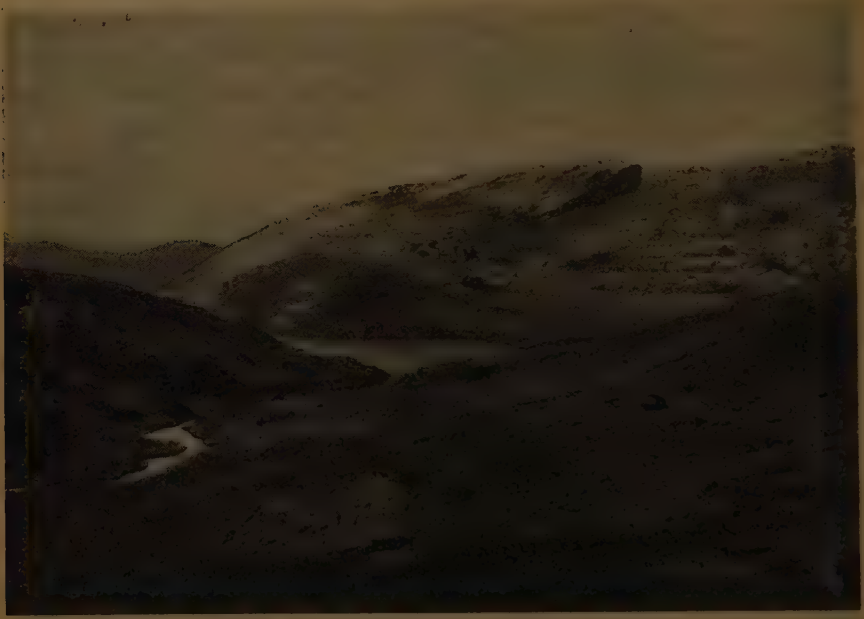


Fig. 2. Contact between micaschist (dark), and quartzitic schist (light) Grønndalen.

THE ROCKS OF THE ANORTHOSITE KINDRED

These rocks constitute the greater part of the field and will, first of all, be dealt with here. The distribution of the various types is seen on the map.

As mentioned above the area was involved in the folding of the Caledonian Mountain Range, and the rocks are clearly marked by this. Only small areas of unaltered igneous rocks are preserved. The basic rocks have been altered into amphibolites and saussurite-gabbro-schist, while the intermediate and acid rocks have become acid biotite-gneisses. The latter are most common.

In the anorthosite massif Valafjellet — Larsfonnfjellet — Brekkenipo the rocks are little influenced by the metamorphism. In the massif and around it older structures are to a great extent preserved. The structures follow the boundaries, and the dip is everywhere towards the massif. The anorthosite consequently lies in a bowl surrounded by acid and intermediate rocks, chiefly mangerites. It is worth noting that the igneous rocks which are least influenced by the metamorphism are found within the anorthosite area or in its immediate vicinity.

In the following will be given a detailed description of the various rock types, their geology and petrography. From the map it can be seen that the unaltered igneous rocks only constitute a small part of the area, but their description will occupy a relatively broad space, because these rocks first of all give the clue to the understanding of the genesis of the rocks. Before proceeding to the description of the rocks, it is necessary to say a few words about their systematic position.

Systematics.

In Goldschmidt's scheme of igneous rocks mangerites and jotunnorites have been placed between monzonite and gabbro. Since this scheme was set up systematics have made great progress, and it may therefore be of interest to try to fit these rocks into a modern system.

In the last years a number of proposals have appeared for quantitative-mineralogical classification of igneous rocks. The best known systems have been set forth by Shand (1927), Johannesen (1931—37), Tröger (1935), and Niggli (1931 and 1935). I have chosen to follow Niggli's system from 1931.

On fig. 3 a Niggli's system is reproduced; in fig. 3 b an attempt has been made to place rocks of the anorthosite kindred in this system. Only the upper half of the scheme is reproduced, the lower half comprising feldspathoid-carrying rocks. The classification is quantitative-mineralogical; a volumetrical analysis, and in some cases a determination of the plagioclase will suffice to place a rock in the system.

The division between gabbro and diorite is fixed at 50% An in the plagioclase. It may be mentioned that both Tröger and Johannesen use the same principle here. Niggli calls the rocks containing 40—60% An in the plagioclase gabbro-diorites.

It is worth noticing that Niggli considers albite up to An_{10} as alkali feldspar. As undoubted plagioclase is considered An_{20} and upwards. The division between the families is fixed at 12.5, 50, and 87.5% plagioclase of total feldspar, and at 12.5 and 65.5% of quartz. Rocks containing more than 65.5% quartz are referred to the silexite family.

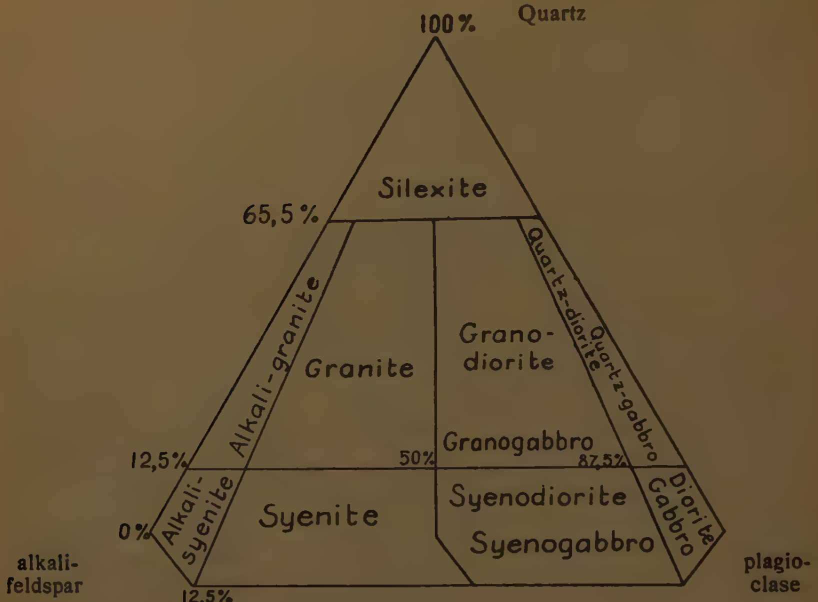


Fig 3 a.

The term silexite has been introduced by W. J. Miller (1919), who has suggested the term for "any body of pure or nearly pure silica of igneous or aqueo-igneous origin which occurs as a dike, segregation mass or inclusion within or without its parent rock. So-called "quartz dikes" are included under silexite."

Below definitions of the most important members of the anorthosite kindred will be stated, and they will be placed in the system where they belong according to their definition. The bracketed figures refer to A. Johannesen's system.

Mangerite-syenite (219) has been defined by N. H. Kolderup in 1920: "Unter dem Namen Mangeritesyenit verstehe ich also ein Gestein mit dem Kieselsäuregehalt des Syenits, und dem für den Mangerit eigentümlichen kalkhaltigen Mikroperthit, samt etwas saurem Plagioklas, mit 10—12% An. Als dunkle Gemengteile finden sich rhombischer Pyroksen, grüne Hornblende oder Biotit, z. T. zwei von diesen Mineralen." The rock consequently falls within the alkali-syenite family.

Mangerite (2210) has been defined by C. F. Kolderup in 1903: "Ein körniges Tiefengestein, das aus einem basischen Mikro-

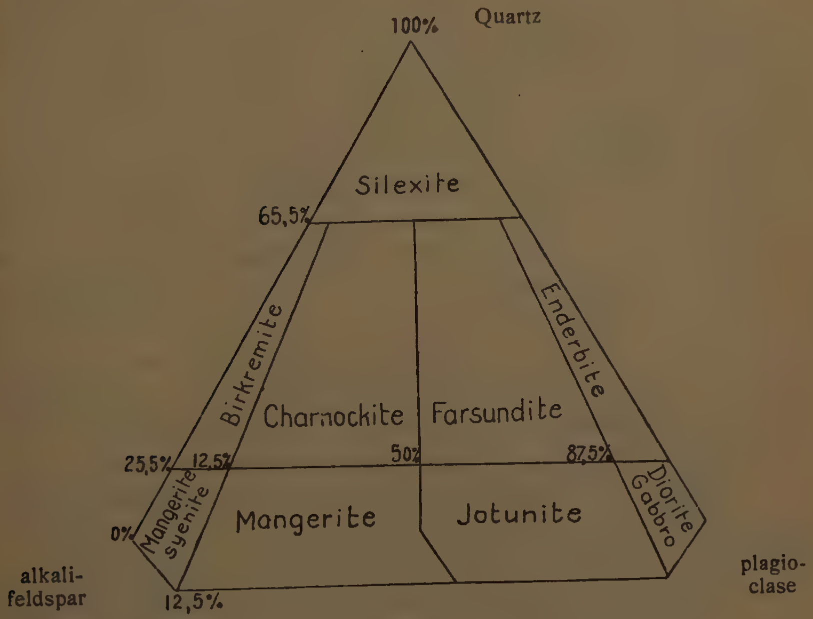


Fig. 3 b.

perthit mit ziemlich großer Menge von Biotit, Hornblende oder Pyroxen besteht, und dessen Zusammensetzung zwischen derjenigen der echten Monzonite und der Gabbros steht."

The mangerite consequently belongs to the syenite family.

Jotun-norite (2211) has been defined by V. M. Goldschmidt in 1916: "Der typische Jotun-Norit besteht aus Plagioklas, Kalifeldspat (meist Mikroperthit), ein wenig Quarz, als helle Minerale, Hypersten, Diallag, etwas Biotit, Erzen, als dunkle Gemengteile, ferner enthält er stets Apatit, sehr häufig Zirkon in auffällig großen, gerundeten Körnern". An % in the plagioclase is set down at 30—35. The jotun-norite consequently belongs to the syenodiorite family. The question then arises whether the term "jotun-norite" can be permitted. As mentioned above both Tröger, Niggli, and Johannesen agree in fixing the division between gabbro and diorite at 50% An in the plagioclase, so that whichever of these systems is followed, the rocks do not fall within the gabbro family.

Another circumstance also makes the term an unhappy one. By norite is understood a gabbro with orthorhombic pyroxene. This generally means a rock in gabbro-facies. In later years it has become more and more usual to let a rock keep its name even though it appears in several facies. E. g. pyroxene-mangerite in gabbro-facies, and amphibole- or biotite-mangerite in amphibolite-facies. When a jotun-norite appears in amphibolite-facies, it has no denomination; to call it an amphibole-jotun-norite is nonsense.

I have here tried to show that the term jotun-norite is no longer convenient, and even though great care should be shown in the changing of rock denominations, I will all the same suggest the adoption of the term jotunite for these rocks. Distinction can then be made between pyroxene-, amphibole and biotite-jotunite in the same way as by the closely related mangerites, and the use of the term "norite" about a rock containing below 50% An in the plagioclase is avoided.

The term anorthosite was employed by T. Hunt (Johannesen 1935) in 1883 about a rock which he believed to be a metamorphic sediment. Now the term has another signification, but its use is not unambiguous. Johannesen restricts the use of the term to a rock of the gabbro family, ɔ: a rock containing more than 50% An in the plagioclase, while Tröger gives the following definition: "Sammelname für Tiefengesteine die fast nur aus Plagioklas bestehen". It is in this sense that the term has been most applied in recent literature. It then comprises rocks both of the gabbro and the diorite families.

Anorthosite-gabbro. Nor in this case has the name been unambiguously employed. Bowen uses it in 1917. He gives no exact definition, but it appears that when the dark minerals in the anorthosite constitute more than 15%, the rock should rather be termed an anorthosite-gabbro. Miller (1919) describes anorthosite-gabbro from Adirondack. He gives no definition of the term either, but he mentions a series of volume analyses. The majority of them have from 15—25% dark minerals. C. F. and N. H. Kolderup (1940) employs the term in the following manner: "A gabbro variety rich in garnet is referred to as anorthosite-gabbro". Here the denomination is used about a rock of the gabbro family where the dark minerals constitute from 15—35%. When the rock has less than 50% An in the plagioclase, anorthosite-diorite might be a suitable term.

Birkremite (215) has been described by C. F. Kolderup in 1896, and defined by him in 1903. The main components are quartz and micro-perthite. When plagioclase is present it is a basic albite. The birkremite then belongs to the alkali granite family.

Charnockite (226) has been defined by Holland in 1900 as the acid member of a definite rock family in India. Corresponding rocks are found in Norway within the anorthosite kindred. Potash feldspar — microcline or microperthite — prevails over the plagioclase. It is an oligoclase with 10—30% An. The charnockite therefore belongs to the granite family.

Farsundite (227). In 1896 C. F. Kolderup described a banatite area near Farsund, but he does not state the quantitative relations between the minerals of the rock. J. H. L. Vogt (1924) distinguishes between farsundite and birkremite in the following way: "In the farsundite an acid plagioclase prevails over potassium feldspar, while in the birkremite the opposite is usually the case". Accordingly the farsundite belongs to the grano-diorite family.

Enderbite (228) has been defined by Tilley in 1936 as a member of the charnockite series. Similar rocks are found in the Bergen field. Feldspar is only plagioclase — an oligoclase — it is frequently antiperthittically developed. The enderbite belongs to the quartzdiorite family.

The rocks of the anorthosite kindred also have representatives within the silexite family; these have no names of their own.

The Anorthosite Massif between Uppheimsdalen and Brekkedalen.

The massif is composed of three mountains traversed by valleys. Between Valafjellet (1162 m) in the south-west and Larsfonnfjellet in the north (1319 m) is Skjulsetdalen, and between Larsfonnfjellet and Brekkenipo in the south-east (1193 m) runs Hodnardalen. The valleys abound in loose material so that exposures can only be found along the rivers. But the mountains themselves are bare, and above 1000 m they are practically void of vegetation and present good conditions for investigations.

Anorthosite is the dominating rock, but gabbro and anorthosite-gabbro are also common, particularly in the boundary areas. Basic segregations of pyroxene, garnet, and hornblende are common; eclogite may also be found.

In Hodnardalen has been segregated a considerable quantity of a rock called by me gabbro "in statu nascendi". Similar rocks have been described by Balk (1931) from the Adirondacks, and the term has been introduced by him.

The structural conditions in and about the massif are especially interesting, and a structural map of the area has therefore been made (Fig. 4). A characteristic of the massif is that it has a foliated boundary zone where the foliation follows the boundary, and the dip is everywhere towards the massif.[1] The anorthosite, then, lies in a bowl surrounded by intermediate and acid rocks, chiefly mangerites. The foliated boundary zone runs along both sides of the boundary, extending upwards into the anorthosite and downwards into the mangerites. These structures show that the anorthosite must have sunk in relation to the surrounding mangerites.

On the north-west and south-east side of the massif these structures run across the Caledonian structures in the gneisses, and may be clearly distinguished from the latter. Undoubted traversing structures have not been found, but from several reasons to which I return later, the structures around the anorthosite massif must be regarded as being older than the Caledonian structures in the gneisses. They must be pre-Caledonian or at any rate early Caledonian. There is much that seems to indicate that they are primary, magmatic.

The two structures are separated by a zone of mylonite, so that the diagrammatic section becomes as follows:

Massive anorthosite.
Foliated anorthosite, and zone of transition to foliated mangerite.
Mylonite.
Mangerite gneiss.

---

[1] As to the use of the terms "foliation" and "schistosity" Harker's definitions from 1932 have been followed.

Fig. 4. Structural map of the anorthosite massif between Brekkedalen and Uppheimsdalen.

1. Contact of the anorthosite. 2. Boundary between the anorthosite and the gabbro "*in statu nascendi*". 3. General strike and dip of the foliation. 4. General strike and dip of the fold axes. 5. Horizontal fold axes. 6. Flow-lines. 7. Vertical flow-lines. 8. Projected strike of flow-lines. 9 Pegmatite. 10. Shear-zones. 11. Fractured zones, mostly breccias and mylonites. 12. General strike and dip of the schistosity.

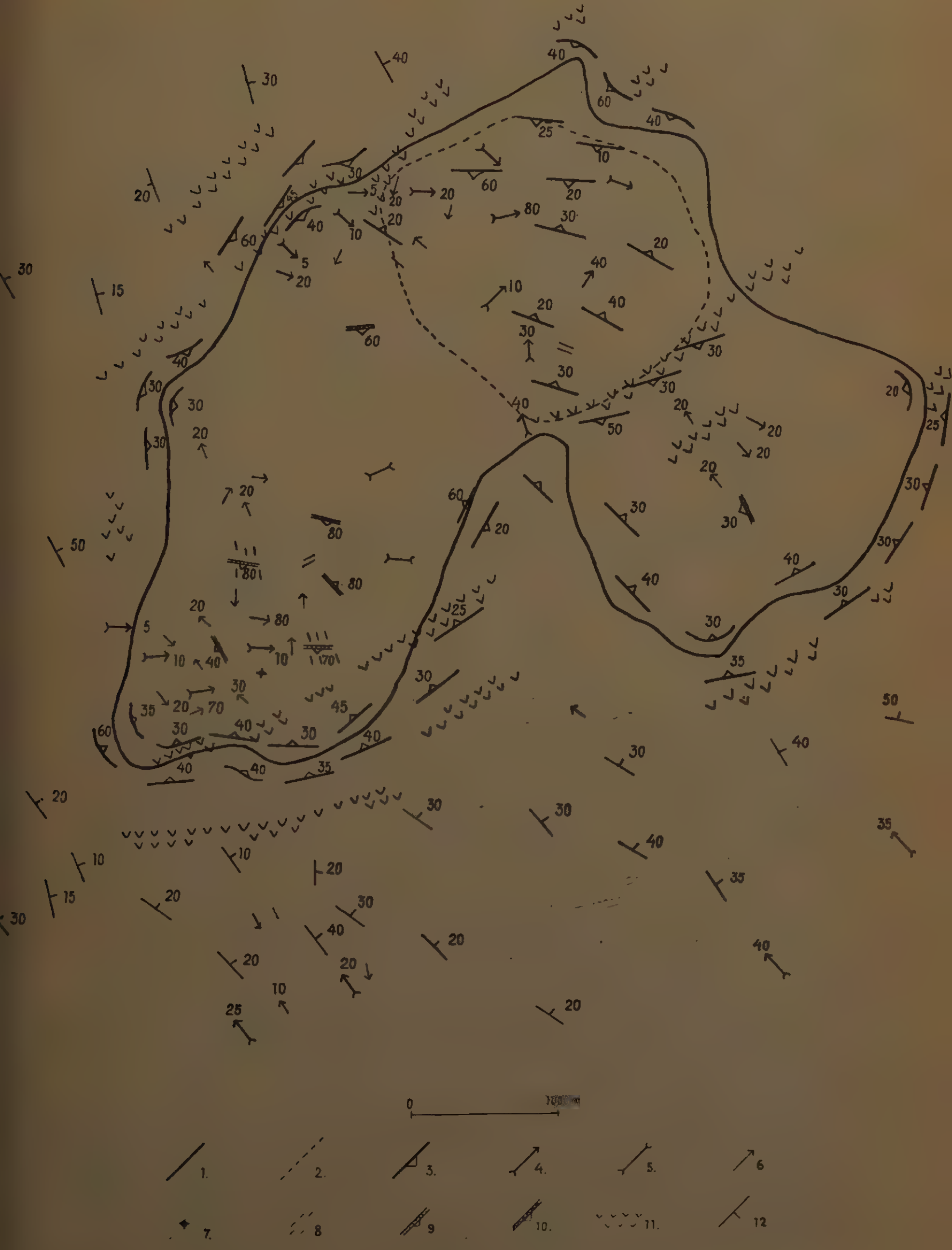


Fig. 4.

The foliated boundary zone around the massif is at the same time a zone of transition to the surrounding acid rocks. There is a gradual transition from anorthosite and anorthosite-gabbro over gabbroic and dioritic rocks to mangerite rocks. This zone extends as a rule over some 100—200 m.

The foliated rocks in the boundary zone contain relatively many dark minerals. Gabbro and anorthosite-gabbro are most common. Inwards into the massif the foliation, as a rule, changes gradually into flow lines, and at the same time the quantity of dark minerals decreases; the rock becomes anorthosite. It is rarely quite structureless. There is a structureless occurence in the top of Brekkenipo and some small areas on the Valafjell. This rock has only an inconsiderable content of dark minerals. It may amount to a few per cent.

Even in the massive anorthosite and in the anorthosite with flow lines narrow foliated zones are common. Their direction is brought out by the structural map. They correspond to what Balk (1931) termed “shear zones”. They are as a rule narrow, a few cm is common, but they may also be somewhat broader. The shear zones consist as a rule of foliated anorthosite. The sketch (Fig. 5) shows such a zone from Brekkenipo.

There is a gradual transition between the massive and the foliated anorthosite. The An % in the massive anorthosite is somewhat higher than in the foliated one; the difference may be about 10—15 per cent. But the feldspar in the foliated anorthosite will always be more or less saussuritized, so that it is difficult to say whether there has originally been any difference.

The shear zones may also consist of more acid material of granitic or mangeritic composition.

The direction of the shear zones varies, which is first of all due to foldings in the anorthosite that have been formed later. The dip is generally steep.

Pure flow lines are common in the anorthosite. They are marked by long spindles of garnet, hornblende, and pyroxene. A series of measurements of the flow lines have been taken, but it is difficult to arrange them into a system. As already mentioned the area is folded. The folds are large and open, and the flow lines take part in the folding. In some places the flow lines may be seen following

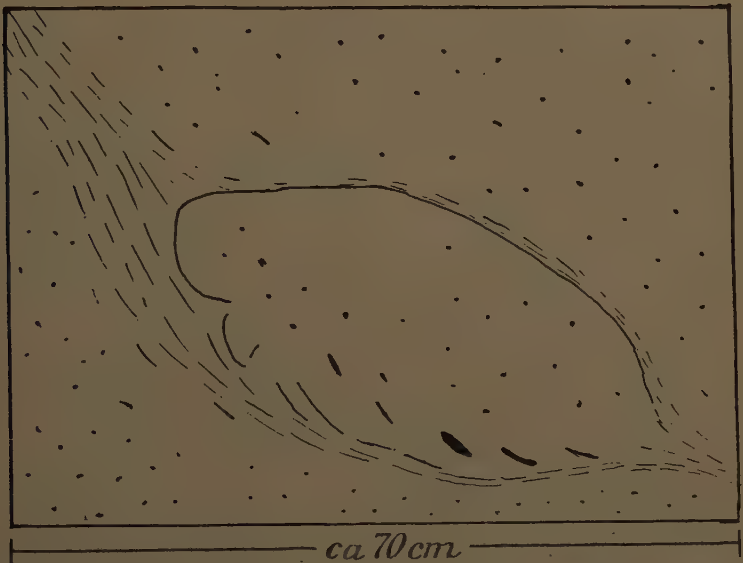


Fig. 5. Shear-zone of foliated anorthosite in massive anorthosite, Brekkenipo.

the foldings all the way round. The dip may be steep, it may rise to 90°, but 30°—40° is more usual.

There is a gradual transition between the flow lines and the narrow bands of dark minerals (Fig. 6). These always run parallel to the flow lines, and are folded together with them. These basic segregations may also form comparatively large round or lens-shaped areas or broad bands. On Valafjellet there are two such large, roundish areas with a diameter of 100—150 m.

The occurrence of silexite is a characteristic feature in the anorthosite massif. It is particularly common in the boundary areas. Here it is foliated, and the foliation is conform with the surrounding rocks. The largest occurrence is on the north-west side of Larsfonnfjell. It has a thickness of some 10 m, and a length of more than 300 m. Also in the interior of the anorthosite massif comparatively small zones of silexite occur. Most extensive is the silexite in the gabbro area in Hodnardalen. Its occurrence here will be described later.

Here I would only mention that the silexite, too, is folded together with the other rocks.

On Valafjell some pegmatite dikes have been observed. They are small, 20—30 cm broad; the largest extends over a couple of hundred meters. These pegmatite dikes cross the flow lines, and must therefore be younger than these. Also in the pegmatite dikes foldings have been observed.

On the structural map a series of foldings are indicated. They have a more regular course than the other structural elements in the massif. On Larsfonnfjell the direction is EW or NW—SE, the measurings varying from 90°—140°. On Valafjell they run EW or NE—SW with variations of 70°—90°. The dip is everywhere nearly horizontal; as a rule it goes inwards to the massif. In Hodnardalen there are a few measurings of axes running NS. Here the foldings are not so common. On Brekkenipo no fold-axes have been measured.

In the gneiss area outside, the fold-axes run parallel to the direction NW—SE with variations of 140°—160°; they have, then, quite another direction. In the anorthosite area the fold-axes show a certain relation to the boundaries, standing approximately vertically to them. If the folds have been formed when the anorthosite was sinking in relation to the surrounding rocks, we should exactly expect the axes to obtain this direction.

A more detailed structural investigation of the area may be able to give more information both about this and about the direction of the flow lines before the foldings.

The structural conditions in the gabbro in Hodnardalen (gabbro "in statu nascendi") require a further description. The rock has a typical granulitic structure with alternating light and dark bands and lenses. There is also a continual alternation between foliated and massive rocks.

As a rule the intermediate and acid rocks are foliated while the basic ones are more massive. The bands and the lenses are drawn out in the strike direction 110°—120°. The dip continues quite up to the level of 1109 m (see map). Here the rock is folded, the dip changes, and becomes S—SW in the slopes down to the Brekkedalen.

The gabbro lies surrounded by anorthosite on all sides. The boundary in the south-east on Brekkenipo is sharp and may be said to have the character of a thrust plane. The gabbro is here dipping under the anorthosite. In Hodnardalen the boundary is covered, and on Larsfonnfjell there is no sharp boundary but a transition. The

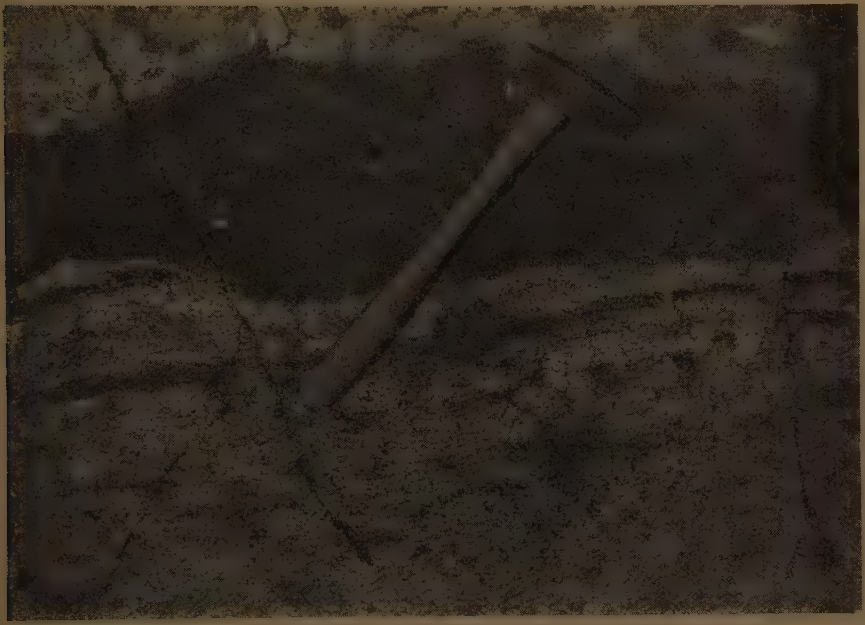


Fig. 6. Dark band and flow-lines in anorthosite, Valafjell.

anorthosite here underlies the gabbro. The latter therefore lies as a lens within the anorthosite.

Finally the most significant features may be summed up: A massif of anorthosite and anorthosite-gabbro rests in a bowl of surrounding mangeritic rocks. These are generally gneissic, but unaltered types are preserved. There is no sharp boundary, but a zone of transition between the anorthosite and the mangerites. This zone of transition is foliated, and the foliation follows the boundary. As a lens in the interior of the massif lies a large area of a banded and schlieren rock, gabbro "in statu nascendi". In the interior of the massif foliated zones of acid mangeritic rocks occur.

The Anorthosite Massif at Slæen-Kamben.

This lies on the outskirts of the field, and has not been examined as thoroughly as the rocks above. It will also prove most natural to examine it in connection with the large anorthosite massif to which it belongs.

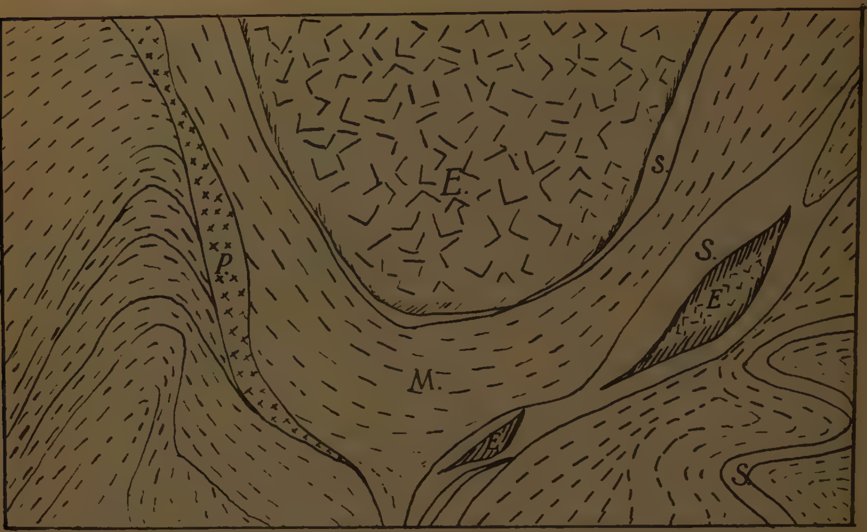


Fig. 7. Road cutting by Tveite. Length of section c. 10 m. E — eclogite, M — mangerite, S — silexite, P — pegmatite.

Petrographically it does not differ from the anorthosite to the west of it, and the samples taken from here will therefore be described together with the others. Only some observations in the field will be briefly mentioned here.

Along the boundary much is covered, but here, too, there appears to be a transition to the surrounding mangerites. In the northwest point of Kamben the anorthosite is saussuritized.

The main road to Stalheim traverses the massif, and here are several fine cuttings, the first one is E of Lunde. The rock here is an anorthosite-gabbro in which garnet dominates among the dark minerals, the plagioclase is saussuritized. Here is a pegmatite dike, and one of silexite. Both are folded.

Farther east there is a magnificient cutting at the farm road to Tveite. The rock here is banded and full of schlieren; it is of the same type as the gabbro of the Hodnardalen. But at the same time it is exceedingly folded, the fold-axes running in several directions.

A sketch shows a section of the cutting and the rock types occurring there (Fig. 7).

The eclogite is traversed by a dense net of joints, but it is otherwise unaltered, and has large, red garnets and green pyroxene. On the left a pegmatite dike is seen which must have been formed later than the folds. It differs from the pegmatite farther west by possessing potash feldspar. As to age the foldings must then lie between the two pegmatite dikes.

On the right a small lens of eclogite is seen, quite surrounded by silexite. The silexite is biotitized at the edges. Similar lenses of eclogite occur also on Larsfonnfjell.

It may also be mentioned that there has been observed a narrow diabase dike intersecting the silexite.

If we follow the road farther east we meet with unaltered white anorthosite, often with a lilac hue. Farther east at Førde the mangerite-gneisses reoccur.

The Interior of the Anorthosite Massif.

In the anorthosite massif it may, as already mentioned, be distinguished between a foliated boundary zone and a more massive area within. In the interior of the massif the mineral assemblage is generally the same in anorthosite, anorthosite-gabbro, gabbro, and the basic segregations. It is only the quantitative relation between the minerals that varies.

The chief minerals are: Plagioclase, pyroxene, hornblende, and garnet. Under the microscope the rocks show signs of strong crushing. They have mortar texture with large grains in a more small-grained crushed ground mass. The ground mass shows, as a rule, no signs of recrystallization. The large grains are generally twisted and fractured; faultings along the joints have been observed. In plagioclase with twin lamellae and in diallage this twisting of the grains is particularly conspicuous (Fig. 8).

The plagioclase is as a rule a labradorite. The following measurements have been taken from samples of unaltered plagioclase:

Anorthosite,	Valafjell	$\alpha_{Na} \sim 1.559$,	ɔ: 61 % An.
—	—	$\alpha_{Na} \leqq 1.559$,	ɔ: 59 »
—	—	$\alpha_{Na} = 1.560$,	ɔ: 63 »
—	Brekkenipo	$\alpha_{Na} \leqq 1.558$,	ɔ: 55 »
—	Ø. Slæen	$\alpha_{Na} \leqq 1.561$,	ɔ: 63 »
Gabbro,	Larsfonnfjell	$\alpha_{Na} \geqq 1.559$,	ɔ: 60 »
—	Slæen	$\alpha_{Na} \leqq 1.559$,	ɔ: 59 »
Eclogite,	Tveite	$\alpha_{Na} \leqq 1.558$,	ɔ: 55 »

As may be seen the variations are not so great. The colour of the labradorite is usually white, but it often has a lilac hue. Under the microscope may then be seen numerous dusty inclusions which are too small to be determined. The colour resembles that of the titanite, and it is not unlikely that they are small inclusions of titanite or some Ti-compound. Twins according to albite and pericline laws are common. Zonar texture has not been observed.

Scapolite occurs, but never in large quantities. As a rule there are small grains occurring on the border between the plagioclase grains. Its quantity is so small that it is difficult to decide whether it is primary or secondary.

The pyroxene is monoclinic; only in a single case orthorhombic pyroxene has been found in the anorthosite massif. The monoclinic pyroxene is macroscopically green with distinct cleavage, and a characteristic metallic glimmering on the cleavage planes. Under the microscope it is colourless to faintly green with marked cleavage after (100). Often there is a segregation of ore grains along these cleavage planes.

In an anorthosite-gabbro from Larsfonnfjell the following measurements of such a pyroxene have been taken:

$2V = 62°$ $(\pm 2°)$ measured with micrometer ocular,
$\beta - \alpha = 0.0065 - 0.007$ measured with Berek's compensator,
Index of refraction $\beta_{Na} = 1.696$.

This corresponds with a diallage.

Hornblende occurs together with pyroxene and garnet. It occurs in independent grains together with pyroxene, and must be assumed to be primary. The colour is green, it may be a brownish green, a greyish green, or a dirty-green, and distinguishes itself from the bluish green hornblende, which among other places is found in the boundary areas.

The pleochroism is distinct, absorption: $\gamma > \beta > \alpha$.

As a type can be chosen a hornblende from a dark band in anorthosite on Larsfonnfjell. The colour is dirty green with distinct pleochroism: $\alpha =$ yellow, $\beta =$ green, $\gamma =$ brownish green. A determination of the indices of refraction gives:

$$\begin{aligned} \gamma_{Na} &\sim 1.688 \ (\pm 0.003) \\ \alpha_{Na} &\sim \underline{1.668} \ (\pm 0.001) \qquad c/\gamma \text{ c. } 20° \\ \gamma - \alpha &= 0.020 \ (\pm 0.004) \end{aligned}$$

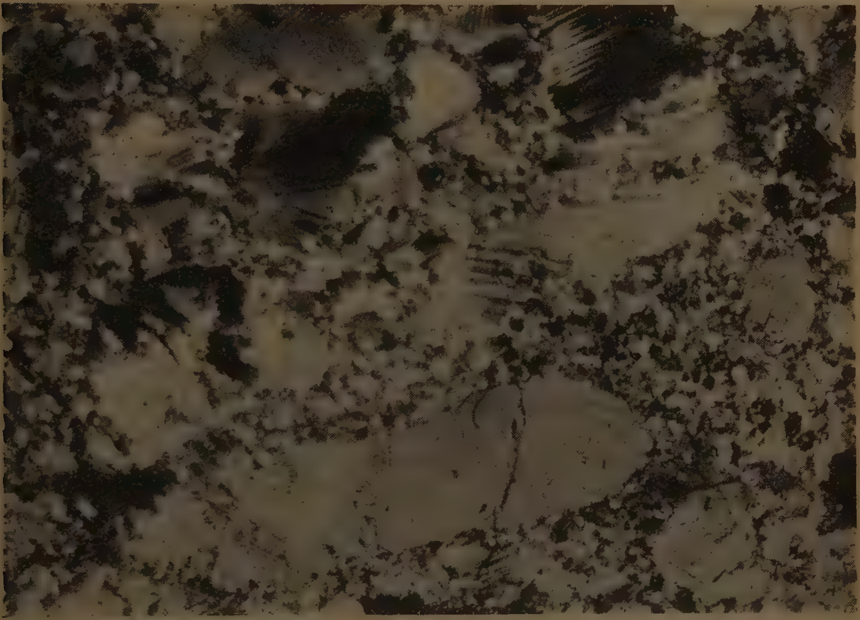


Fig. 8. Anorthosite, Larsfonnfjell. X 10.

According to Kunitz' tables (1930) this corresponds to a basic syntagmatite-hornblende with small Ti-content and 28% Fe-compound.

A brownish grey hornblende from a gabbro from the Larsfonnfjell has somewhat lower index of refraction: $\alpha_{Na} \sim 1.666$. c/γ c. 14°.

Garnet is very common; in the anorthosite-gabbros it may constitute more than 20 vol.%. Macroscopically it is reddish brown, in thin sections pink. The size varies from a few mm up to several cm in diameter. It is seldom idiomorphic, and occurs in most cases in rounded-off grains. Outside the anorthosite area garnet is seldom found; it is found in a few occurrences of gabbro, but in the intermediate and acid members and in the gneisses it is scarce.

As mentioned above the structure of the anorthosite shows signs of strong crushing. Also the garnet is crushed; it is traversed by a series of fissures, which are as a rule filled by a light green biotite or hornblende. A marginal zone of biotite and hornblende around the garnet is also common. These fissure-formations with biotite and hornblende are particularly developed in the boundary areas. In less altered types in the interior reaction rims between garnet-plagioclase and garnet-pyroxene can be observed. But the zones are small and poorly developed, and a determination of the minerals is not possible. The table below shows the composition of the garnet.

No.	Rock	Locality	Index of refraction in Na-light	Composition
1	Eclogite	Tveite	1.753	$Fe_{31}Mg_{56}Ca_{13}$
2	Eclogite	Larsfonnfjell	1.753	$Fe_{31}Mg_{56}Ca_{13}$
3	Pyroxenite	Tveite	1.755	$Fe_{32}Mg_{44}Ca_{24}$
4	Pyroxenite	Larsfonnfjell	1.765	$Fe_{38}Mg_{44}Ca_{18}$
5	Gabbro	Larsfonnfjell	1.753	$Fe_{31}Mg_{46}Ca_{24}$
6	Gabbro	Slæen	1.757	$Fe_{32}Mg_{42}Ca_{24}$
7	An-gabbro	Hodnarelven	1.763	$Fe_{37}Mg_{39}Ca_{24}$
8	Anorthosite	Valafjell	1.757	$Fe_{33}Mg_{43}Ca_{24}$
9	Gabbro	Hodnarelven	1.791	$Fe_{58}Mg_{27}Ca_{15}$
10	Gabbro	Tveite	1.791	$Fe_{58}Mg_{27}Ca_{15}$

By the determination Eskola's tables have been used (Eskola 1921). The exactitude of the measurings varies from ± 0.002 to ± 0.004. The Mn-content has not been examined. The Ca-content is taken from the table in all cases except for No. 4. Eskola has an analysis of "garnet from labradorite-rock Aurlandsfjord". The index of refraction is here the same as that of the pyroxene from the Larsfonnfjell. They have therefore been given the same Ca-content. It is probable that analyses would show a somewhat lower Ca-content for several of the other samples also.

It is seen from the table that the variation of composition is slight, only the two last samples show any note-worthy deviation. These two samples will be further treated later, but otherwise all the garnets are rich in pyrope; they are typical eclogite-garnets. The conditions for the formation of such minerals are high pressure and high temperature. They belong to the eclogite- and the granulite-facies.

The manner in which the garnet occurs shows that it must have been formed at an early stage, presumably contemporanously with the other minerals within the anorthosite massif. It must at any rate have been formed before the crushing, being itself crushed and the grains rounded-off. Another indication of early origin is the fact that the garnet forms the flow lines of the anorthosite together with pyroxene and hornblende.

Biotite is found as a subordinate mineral in the interior of the anorthosite massif. It occurs together with the other dark minerals, or in rosettes around the ore. It is brown with strong pleochroism. It is more frequent in the border areas. Both green and brown biotite are found; the green is always found in the vicinity of the garnet, most often on joints in the latter.

Epidote mineral occurs, but it is most common in the border areas. In accessory amounts are found: Ore (ilmenite) and zircon.

The Boundary Areas of the Anorthosite.

As mentioned above there is a clear difference between the interior and the boundary areas of the anorthosite. In the interior the rocks are predominately massive, and anorthosite is the dominating type. In the boundary areas the rocks are distinctly foliated, and they contain at the same time considerably more dark minerals. Anorthosite-gabbro and gabbro predominate. The foliation follows the boundary, and there is as a rule a gradual transition to the massive types in the interior.

Under the microscope the difference between the two types appears more clearly. The structure of the interior is, as mentioned above, considerably crushed. In the boundary areas there has to a great extent been a recrystallization of the plagioclase. This appears clearly from a series of samples taken from the boundary area on the south-west slope of the Larsfonnfjell. The samples have been taken at intervals of about 50 m.

No.	An % in the plagioclase	Index of refraction in Na-light	Rock
1	28	$\alpha_{Na} \sim 1.543$	Diorite
2	46	$\alpha'_{Na} \sim 1.552$	An.-gabbro
3	50	$\alpha_{Na} \leq 1.554$	»
4	26	$\alpha'_{Na} \sim 1.541$	Mangerite
5	59	$\alpha'_{Na} \sim 1.559$	An-gabbro
6	55	$\alpha_{Na} \sim 1.557$	»
7	63	$\alpha_{Na} \sim 1.560$	Anorthosite

The accuracy of the measurements varies from ± 0.001 to ± 0.003. Where α is primed, the measurements have been taken on the cleavage planes, and the table of Rosenbusch (1927) for index of refraction on the cleavage planes has been used. In the other cases the usual table for index of refraction has been used, also according to Rosenbusch.

The first sample is from a zone of crushing under the anorthosite. In the next two the plagioclase is completely recrystallized, the rock showing no signs of crushing. The fourth is strongly pressed mangeritic zone within the anorthosite. The next two display partial recrystallization. Larger grains that show signs of crushing lie in the more fine-grained ground mass where the recrystallization has been considerable. The seventh finally shows pure mortar texture without any trace of recrystallization, such as is usual for the anorthosite in the interior of the massif (Fig. 8).

In the samples having two generations of plagioclase grains, the An % is the same in either grain. This has been controlled by investigating whether there is any difference of index of refraction in two equally orientated neighbouring grains. It is worth noticing that the recrystallized grains display no signs of crushing. That implies that the crushing in the interior of the anorthosite area must be older than the recrystallization in the boundary areas.

The series also show nicely how the An % in the plagioclase increases from the boundary inwards. No. 4 is, as mentioned, a mangeritic zone in the anorthosite.

Other measurements from the boundary area show that the plagioclase has an An % of 40—50. The links of transition to the underlying mangerites are somewhat lower in An.

The decrease of the An % in the plagioclase is followed by an increase of the quantity of epidote. The volume analysis of No. 7 (see p. 166) shows 63% An in the plagioclase and 1.6 vol.% epidote. The volume analysis of No. 3, which is a typical anorthosite-gabbro from the boundary area, has 50% An in the plagioclase and 3.51 vol.% epidote-mineral.

Other samples also show that there is a distinct relation between the content of lime in plagioclase and the quantity of epidote. If the equilibrium of epidote and plagioclase is used as a geological P-T indicator, this shows that the temperature during the consolid-

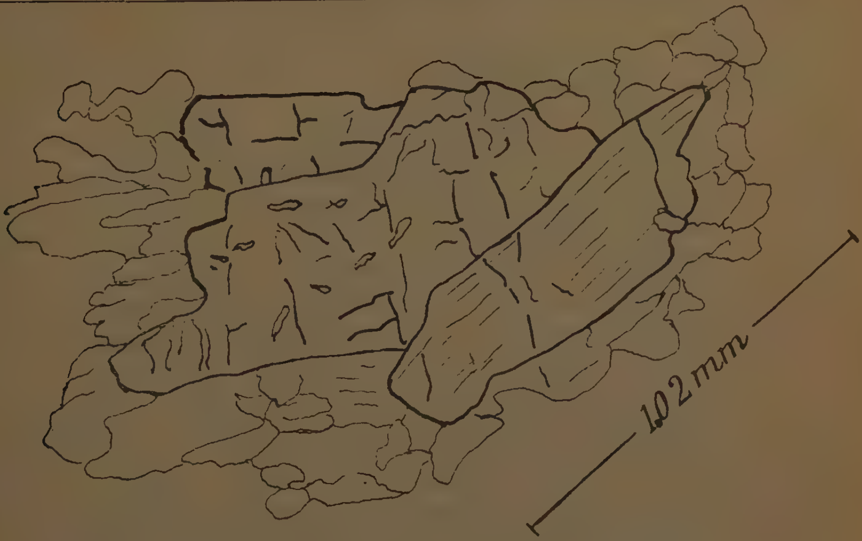


Fig. 9. Clino-zoisite in anorthosite-gabbro surrounded by hornblende and biotite.

ation has been higher in the interior than in the boundary areas. In the interior the temperature has been so high that a plagioclase with some 60 An % has been stable. In the boundary areas the temperature must have been lower, a more acid plagioclase has been stable, and the additional Ca has been absorbed by the epidote-mineral. As mentioned above idiomorphic epidote occurs together with a plagioclase as basic as An_{68}. According to Ramberg (1943) this should correspond to a temperature of more than 500°.

The epidote-mineral may be zoisite, but most frequently it belongs to the clino-zoisite—epidote series. The zoisite has parallel extinction and has abnormal blue interference colours. All epidote-minerals occur together with hornblende, biotite, and garnet, most frequently in intimate mixtures. They almost show clear crystal outlines, and are found in long columnar grains (Fig. 9).

The epidote here occurs in such a manner that it must be regarded as equal to the other minerals.

Hornblende is the dominant dark mineral in the boundary areas. Pyroxene is found; in some thin sections it can be clearly seen that the hornblende has been formed at the expence of the pyroxene. But most frequently there has been a complete recrystallisation, so that the hornblende is alone prevailing.

Measurements from an anorthosite-gabbro from Larsfonnfjell show the following figures:

$$\begin{aligned} \gamma_{Na}\ & 1.683\ (\pm 0.002)\quad c/\gamma \sim 22^\circ \\ \alpha_{Na}\ & \underline{1.663}\ (\pm 0.003) \\ \gamma - \alpha = & 0.020\ (\pm 0.005) \end{aligned}$$

Absorption: $\gamma > \beta > \alpha$. Pleochroism: $\gamma =$ bluish green, $\beta =$ green, $\alpha =$ yellowish green. According to Kunitz' tables this corresponds to a syntagmatite-hornblende with 54 % Fe compound. It is often partly altered to biotite.

Biotite. In the boundary areas both brown and green biotite are generally found; both may occur in the same thin section. The brown one has strong pleochroism: $\gamma = \beta =$ dark brown, $\alpha =$ yellow. It is typical as an alteration product of the hornblende. The green one is particularly light, and the pleochroism is not pronounced: $\gamma = \beta =$ light green, $\alpha =$ light yellowish green. It is always found in the neighbourhood of the garnet, often on joints in it. It is under transformation to chlorite, but has preserved the interference colours of the biotite.

Garnet is also found in the boundary areas. It is the same garnet rich in pyrope that is found in the interior, but in the boundary areas it would seem to be unstable. It is intersected by a number of fissures, filled with hornblende, biotite, and an epidote-mineral. These fissure minerals increase at the expense of the garnet. In some samples only quite narrow streaks of the garnet are left. (Fig. 10.)

The dark minerals are assembled in lumps or bands. The characteristic association is an intimate mixture of garnet, hornblende, biotite, and an epidote-mineral.

Above a general survey of the distribution of the various rocks within the anorthosite massif has been given. Below a more detailed description of the different types will follow.

Ultra-basic Rocks.

Eclogite never occurs in large quantities, but in small bands it is quite common. The largest observed occurrence is at a crossroad at Tveite (see p. 148). Here is a lens with a diameter of a couple

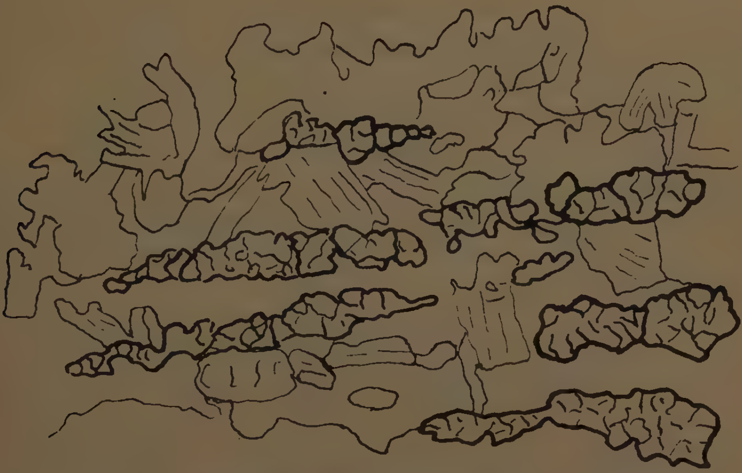


Fig. 10. Micro-drawing from anorthosite-gabbro, Larsfonnfjell.
It shows narrow streaks of garnet surrounded by hornblende, biotite and epidote.

of meters. The rock is greatly crevassed. Macroscopically garnet and a dark, green mineral are seen. The structure is eugranitic. Under the microscope a net of parallel joints or weak zones is seen. The mineral grains show incipient alteration along these zones.

The chief minerals are: Garnet, olivine, pyroxene, and hornblende. Minor constituents: Plagioclase and ore.

The garnet is isotropic. The colour is pink to faintly brownish. It shows partly idiomorphic outline, and has inclusions of pyroxene and hornblende.

Sp. g = 3.77 determined with Clericis' liquid. Index of refraction determined with immersion liquids, $n_{Na} \sim 1.753$ ($\pm$ 0.003). According to Eskola (1921) this gives the following composition: $Fe_{31}Mg_{56}Ca_{13}$, i. e. a typical eclogite- garnet.

The pyroxene is colourless and shows no pleochroism. Indices of refraction:

$$\begin{aligned} \gamma_{Na} &\sim 1.707 \;(\pm\, 0.003) \\ \alpha_{Na} &\sim \underline{1.681} \;(\pm\, 0.001) \\ \gamma - \alpha &\sim 0.026 \;(\pm\, 0.004) \quad c/\gamma\ 40^\circ\ 2V \sim 65^\circ \end{aligned}$$

In comparison it may be mentioned that Eskola has found the following values for a pyroxene from an eclogite from Silden, Nordfjord (Eskola 1921): $\gamma - \alpha = 0.24$ $2V = 67^\circ$ c/γ c. 42°. As will be seen a very good accordance.

The olivine is also colourless, has irregular cleavages, high interference colours, and paralel extinction. It is comparatively fresh, a little altered on the fracture lines and along the fissures. Here small ore grains have often been segregated.

The hornblende is brown with strong pleochroism: $\gamma =$ chocolate brown, $\beta =$ light brown, $\alpha =$ light yellow. It has two marked cleavage directions; in basal section the angle between them is 124°.

Indices of refraction:

$$\begin{aligned} \gamma_{Na} &\sim 1.676\ (\pm\ 0.002) \\ \alpha_{Na} &\sim \underline{1.654\ (\pm\ 0.002)} \\ \gamma - \alpha &\sim 0.022\ (\pm\ 0.004)\ \ c/\gamma \sim 16^\circ,\ \text{dispersion: } \varrho < v \end{aligned}$$

According to Kunitz' tables this corresponds to a syntagmatitic hornblende with c. 40 % Fe.

The plagioclase is unaltered. It occurs in small roundish grains among the other minerals. It has twin lamellae according to the albite and pericline laws.

Index of refraction: $1.559 > \alpha_{Na} \geqq 1.558$, in section $\perp \gamma$, ɔ: 59 (± 1) %. An.

Where the weak zones mentioned above intersect the plagioclase grains, small grains of sericite, possibly also of epidote have been formed. There is not much plagioclase; it may be estimated to below 10 %. Reaction rims appear between the various minerals, especially between hornblende and garnet, but they are small and badly developed, so that it is not possible to determine the minerals they contain.

A thin section of an eclogite lens from the Larsfonnfjell shows exactly the same mineral assemblage. The only difference is that the rock is here more crushed and somewhat more fine-grained. This eclogite lens lies surrounded by silexite. Around the eclogite lens at Tveite there also runs a narrow dike of silexite.

Garnet-pyroxene-hornblende Rocks. In the anorthosite area dark segregations abound, chiefly consisting of garnet, pyroxene, and hornblende. The garnet may be concentrated as a marginal zone around the pyroxene and the hornblende, but there may also be small aggregates of garnet, evenly dispersed among the pyroxene and the hornblende, or bands of garnet may alternate with bands of pyroxene and hornblende.

Such a garnet-pyroxene-hornblende rock from Larsfonnfjell will be described. Macroscopically bands of garnet and a little plagioclase alternate with bands of a dark mineral where distinct cleavage planes are seen. Under the microscope it shows the following mineral assemblage: Garnet, pyroxene, hornblende, and plagioclase. The texture is typical mortar texture. Great grains of pyroxene and hornblende lie in an exceedingly crushed ground mass.

The pyroxene is colourless, and has a pronounced cleavage parallel (100). Ore has often been segregated along the fissures. The grains generally are twisted and bent. Index of refraction: $\alpha_{Na} > 1.692$ ($\pm$ 0.001) c/γ 44°. This corresponds with a diallage.

The hornblende has distinct pleochroism: $\gamma =$ greyish green, $\beta =$ brownish green, $\alpha =$ yellowish green. c/γ c. 20°. Indices of refraction:

$$\begin{aligned} \gamma_{Na} &\sim 1.688 \ (\pm\ 0.003) \\ \alpha_{Na} &\sim \underline{1.668 \ (\pm\ 0.002)} \\ \gamma - \alpha &\sim 0.020 \ (\pm\ 0.005) \end{aligned}$$

According to Kunitz (1930) this corresponds to a basaltic syntagmatite-hornblende, low in Ti with c. 28 % Fe compound. It occurs in independent grains together with the pyroxene and must be taken as primary.

The garnet is pink of colour. It is isotropic. It is excessively crushed, and does not show idiomorphic outline.

Index of refraction: $1.770 > n_{Na} > 1.760$. According to Eskola (1921) this gives the following composition: $Fe_{38}Mg_{44}Ca_{18}$. The Ca content is here put equal to 18, after an analysis by Eskola (1921, p. 109) of a garnet from the labradorite rock Aurlandsfjord. It has an index of refraction of 1.7658.

Plagioclase is found as a minor constituent. It is unaltered and it has twins according to the albite and pericline laws. This rock distinguishes itself from the eclogite by the fact that the garnet contains more iron, and that the pyroxene is a typical diallage.

A similar rock is found at Tveite, south of the main road. It has somewhat more plagioclase, and forms in so far a transition to gabbro. It has the following mineral assemblage: Pyroxene is monoclinic. It is colourless and excessively crushed. $c/\gamma \sim 40°$. 2V c. 60°. It is partly altered to bluish green hornblende. The hornblende shows pleocroism: $\gamma =$ bluish green, $\beta =$ green, $\alpha =$ yellowish green. c/γ c. 18°. The garnet is rich in pyrope. Index of refraction: $n_{Na} = 1.755$ ($\pm$ 0.001). Of light minerals there is besides plagioclase a little scapolite. Secondarily a little clinozoisite has been formed. Here are reaction rims between garnet—hornblende, garnet—pyroxene, pyroxene—plagioclase. Most common is a zone of bluish green hornblende around the garnet. The zone is fibrous, and the fibres stand vertically to the garnet.

Basic Rocks.

Gabbro. As mentioned areas of gabbro are common within the anorthosite area. There are all transitions from the basic segregations to gabbro and further to anorthosite-gabbro.

As a type a garnet-hornblende-gabbro from Slæen will be described. It is medium-grained, fresh, with eugranitic structure. Macroscopically plagioclase, garnet, and hornblende are seen Under the microscope biotite and scapolite are seen in addition. The plagioclase is a labradorite. Index of refraction: $\alpha_{Na} \leq 1.559$, ɔ: 60 % An. The hornblende is brownish green as is usual within the anorthosite area, c/γ c. 20°. The biotite is brown; the garnet is a usual garnet rich in pyrope. The scapolite is found in small grains, most frequently surrounded by plagioclase grains, or on the border between the latter and garnet. A volumetrical analysis shows the following figures:

Plagioclase	43.2 vol.%	
Hornblende	28.2 »	
Garnet	27.0 »	% light minerals: 43.4 (scapolite is taken as a light mineral).
Biotite	1.4 »	
Scapolite	0.2 »	

From Larsfonnfjell has been examined a gabbro containing pyroxene, a diallage, alongside the hornblende. The plagioclase is somewhat basic. Index of refraction: $\alpha_{Na} = 1.559$ in section $\perp \gamma$ ɔ: 61 % An. Besides the mineral assemblage is the same.

Small areas of gabbro occur also outside the anorthosite area. The largest is in Uppheimsdalen just north of the mica schist. Unfortunately the ground is so covered here that it is difficult to state anything about the structural conditions. Only a few crags project, and the structure in these is eugranitic. The area is lens-shaped, with its longest axis c. 500 m; its direction is NE—SW. In a place on the boundary areas of the mangerite-gneisses dioritic links of transition have been found, so that there is some evidence of a gradual transition to the more basic rocks around. In appearence it is dark, medium-grained and with eugranitic structure.

Microscopical investigations:

A volumetrical analysis shows the following figures:

Hornblende	45.5	vol.%	
Plagioclase	42.7	»	
Chlorite	7.0	»	% dark minerals: 57.3.
Garnet	2.3	»	An % in the plag.: 44.5.
Ore	1.7	»	The plagioclase is not fresh.
Apatite	0.8	»	

The hornblende is greyish brown with pronounced pleochroism: $\gamma =$ greyish brown, $\beta =$ yellowish brown, $\alpha =$ yellowish grey. Absorption: $\gamma > \beta > \alpha$.

$$\begin{array}{rl} \gamma_{Na} \sim & 1.676 \ (\pm 0.002) \\ \alpha_{Na} \sim & \underline{1.656 \ (\pm 0.002)} \\ \gamma - \alpha \sim & 0.020 \ (\pm 0.004) \quad 2V = \text{c. } 85^\circ \quad c/\gamma \text{ c. } 19^\circ \end{array}$$

According to Kunitz' tables (1930) this corresponds to a syntagmatite hornblende with c. 42 % Fe compound.

The plagioclase is somewhat altered. Some grains are filled with a dense mass of sericite, other grains are filled in the middle and have a fresh zone at the edges, while others again lack sericite completely. Index of refraction: $\alpha_{Na} = 1.552$ in section $\perp \gamma$ ɔ: 44.5 % An.

Because the plagioclase is not fresh, I have considered the rock as belonging to the gabbro family, but it stands on the border of the diorite family.

The garnet is pink-coloured and isotropic. It is traversed by a dense net of joints filled with chlorite. Because of this I have not been able to obtain satisfactory measurements of the index of refraction.

The chlorite has generally inclusions of small ore grains. It has abnormal bluish interference colours, and consequently is a penninite.

The apatite occurs in small rounded-off grains.

The other gabbro-zones are so small that they have not been indicated on the map. At Ettesvoll, north-east of lake Uppheimsvatn there is a small occurrence of an uralite-gabbro, which will be further mentioned. The mineral assemblage is: Plagioclase, apatite, pyroxene with a marginal zone of hornblende, biotite, garnet, and ore. The plagioclase is twinned according to the albite and pericline laws; some grains have zonal texture. Index of refraction: $\alpha_{Na} = 1.559$ in section $\perp$ γ ɔ: 61 % An. The apatite is abundant; it is partly idiomorphic.

Hornblende and pyroxene always occur together; as a rule there is a core of pyroxene which is more or less altered; it is surrounded by a coating of hornblende. The pyroxene is monocline, c/γ 42°. It is colourless and has a good cleavage parallel (100).

The hornblende is pale green with slight pleochroism. Index of refraction: $\gamma_{Na} \leqq 1.673$. The determination is difficult because the coating of the hornblende is composed of very small grains with various extinctions. The index of refraction lies lower than for the hornblendes hitherto mentioned. According to Kunitz' tables this corresponds to a syntagmatite-hornblende with c. 38 % Fe compound.

Often there are two distinct marginal zones round the pyroxene, the outer one being then colourless or having a faint greenish tinge. It may also happen that the whole grain is filled with small needles of hornblende, so that only the outer form of the pyroxene is preserved. These marginal zones are considered to be late magmatic formations.

The occurrence of the garnet has a different character from those treated above. It has a core of ore, and outside this there are ore inclusions. It usually displays idiomorphic outlines. All this is evidence of the late formation of the garnet.

A similar garnet occurs somewhat farther east, in a gabbro at Tveite. This gabbro is strongly pressed, and possesses the Caledonian schistosity. The mineral assemblage is the same as that of the gabbro at Ettesvoll, apart from some quartz that has been formed later. Here the garnet is clearly secondary with spiral inclusions of ore, mica, and plagioclase, demonstrating that the garnet has rotated during its growth. The index of refraction of this garnet is: $n_{Na} = 1.791$, and the composition then becoming: (see No. 10, p. 152) $Fe_{58}Mg_{27}Ca_{15}$. This is one of the garnets of the table which is distinguished from the others by containing more iron. It is worth noting that the garnet here also disinguishes itself from the rest by its mode of occurrence.

These are the only two cases where a growth of the garnet has been observed. The two samples mentioned are found outside the anorthosite area proper.

Gabbro "in statu nascendi". The rock lies as a great lens in the interior of the anorthosite area. As mentioned above it has granulite structure. Dark and light bands and lenses alternate. Among these bands the following rock types have been determined: Eclogite, gabbro, anorthosite-gabbro, mangerite, granitic rocks, silexite, and pegmatite, i. e. the greater part of the rocks of the anorthosite kindred that are otherwise common in the field. By collecting more samples representatives of the other members of the kindred would probably also be found.

These rocks are not distinguished from those otherwise occurring in the field, and there is no reason why they should be separately mentioned. The only difference is their being usually somewhat more strongly pressed and crushed than the rest (Fig. 11).

As mentioned above Balk (1931) has described similar rocks from the Adirondacks, characterizing them in the following way: "In many gabbroic rocks the dark silicates are not concentrated, but mixed up with appreciable quantities of anorthosite or syenite. Such rocks are believed to represent the incipient stages in the magma, and when the final consolidation overcame the rock, the 'end stage', that of a normal gabbro, had not yet been attained.

Such accumulations of dark silicates the writer proposes to call gabbros *"in statu nascendi"*."

Besides these schlieren rocks there is an occurrence of an eugranitic rock which is unusually hard. Macroscopically garnet is seen and a dark greenish mineral in a dense ground mass of greyish feldspar. Under the miroscope it is seen that the rock has been exposed to very strong crushing. It has the following mineral assemblage: Plagioclase, $\alpha_{Na} \sim 1.552 \ (\pm 0.001) \perp \gamma$, i. e. c. 45 % An. It is comparatively fresh. Potash feldspar abounds. The ratio plagioclase—potash feldspar is estimated at 50 : 50. The potash feldspar is chiefly a microperthite with a few great round spindles.

Monoclinic pyroxene, the colour of which is light green. It cleaves easily after (100). c/γ c. 40°. This corresponds with a diallage. It is excessively crushed.

Garnet occurs in considerable quantity in large rounded grains. Index of refraction: $n = 1.791$ (see No. 9, p. 152), and the composition accordingly: $Fe_{58}Mg_{27}Ca_{15}$. This is then the second of the two garnets showing deviating composition. There is here no evidence of the garnet being secondary.

Titanite is exceptionally abundant, the content may be estimated at between 10 and 20 vol.%. The mineral occurs in large rounded grains. One of the largest measures 0.78 × 0.42 mm. Polysynthetic twins are common. It has strong dispersion of the optical axes, $\varrho > \nu$.

The rock contains a little ore, presumably ilmenite. Secondarily have been formed: Calcite, muscovite, and a little epidote.

Where this rock should be placed systematically is not easy to say. The ground mass is so broken-up that it is impossible to get a satisfactory volumetrical analysis of it. It probably belongs to the syenodiorite family.

Anorthosite-gabbro, as mentioned above, is found primarily in the boundary areas of the anorthosite massif. In the field it is often difficult to decide whether a rock is a gabbro or an anorthosite-gabbro, but during the microscopic examination it becomes apparent that the majority of the foliated boundary rocks lie on the border between gabbro and anorthosite-gabbro. Here some volume analyses of typical rocks from the boundary region will be given. It was endeavoured to obtain a collection as representative as possible. This is not always easy, the rock being banded, and the dark minerals being often assembled in lumps.

A sample from the Larsfonnfjell (No. 5 in the table p. 153):

Plagioclase	59.7	vol.%	59 % An.
Pyroxene	6.8	»	
Garnet	21.5	»	% dark minerals: 40.3.
Hornblende and biotite ...	12.0	»	

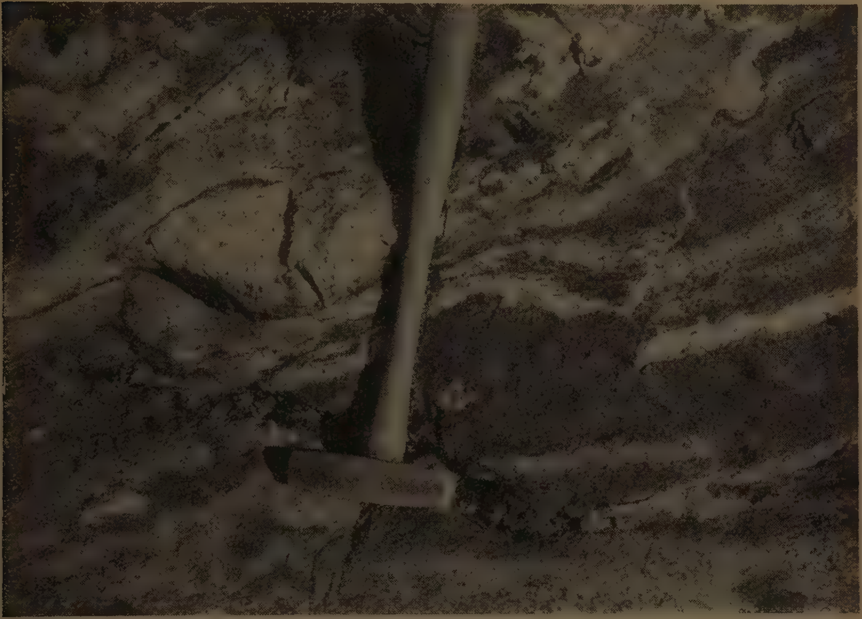


Fig. 11. Gabbro *"in statu nascendi"*, Hodnardalen.

The rock is accordingly a gabbro, but it is on the transition to an anorthosite-gabbro. The pyroxene is a diallage. The hornblende is usually bluish green; it has partly been altered to biotite, and occurs in intimate mixture with it. In the integration they are therefore taken as a whole. The garnet is a common eclogite-garnet rich in pyrope.

No. 3 in the table p. 153 has the following mineral composition:

Plagioclase	64.32	vol.%
Hornblende	29.95	»
Clino-zoisite ...	3.15	»
Biotite	1.42	»
Titanite	0.48	»
Muscovite	0.68	»

Index of refraction: $\alpha_{Na} \sim 1.554$, on the cleavage planes. ɔ: 50 % An. % dark minerals 35.68; if clino-zoisite is taken as a light mineral, the percentage of dark minerals becomes 32.53.

The rock is here a gabbro or an anorthosite-gabbro according as the clino-zoisite is considered a dark or a light mineral. The minerals are otherwise the same as those described above from the boundary zones of the anorthosite. It should be noticed here that the clino-zoisite is found in independent, separate grains of equal rank with the other minerals. It is commonly found in columnar grains which are often very large. Two of the largest grains have been measured. They are 1.02 mm × 0.3 mm, and 0.96 mm × 0.6 mm, respectively (Fig. 9).

It is worth noting that epidote-minerals are lacking in the sample containing pyroxene. Epidote is otherwise common in the border region, occurring as a rule together with hornblende and biotite.

Anorthosite. The main features of the anorthosite have been described above (p. 149). Some additional remarks will be made here. A volume analysis of an anorthosite from the Larsfonnfjell has been performed. Macroscopically it is eugranitic with white plagioclase, and poor in dark minerals.

Plagioclase	92.0	vol.%	$\alpha_{Na} \sim 1.560$ ɔ: 63 % An.
Garnet	2.6	»	% dark minerals: 8.
Hornblende	0.9	»	
Biotite	2.8	»	
Epidote-mineral	1.6	»	
Scapolite	0.1	»	

A notable feature of this rock is the fact that the epidote-mineral is idiomorphic. The plagioclase is fresh; the structure very brecciated (Fig. 8), and there has been no recrystallization in the ground mass. The epidote cannot therefore have been formed later at the expense of the plagioclase, but must be of equal rank with this and the other minerals. These are the same minerals as those which are otherwise found in the anorthosite area. The garnet is rich in pyrope, the hornblende is a bluish green of the common type, the biotite is brown, and the epidote-mineral belongs to the clino-zoisite—epidote series.

It has been mentioned above that zones of foliated anorthosite occur in the massive one. On the Valafjell samples of such a foliated zone and of the massive rock near by have been taken. The index of refraction of the plagioclase in these samples was examined.

a) massive anorthosite: $\alpha_{Na} \leqq 1.559 \perp \gamma$ ɔ: 59 % An.
b) foliated anorthosite: $\alpha_{Na} \leqq 1.553 \perp \gamma$ ɔ: 45 % An.

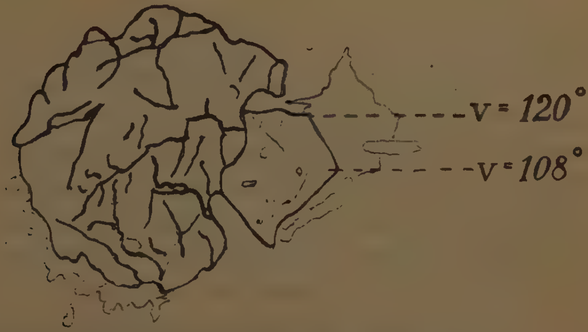


Fig. 12. Idiomorphic epidote in anorthosite, Larsfonnfjell.

The foliated anorthosite is strongly saussuritized, a feature common to all the foliated zones. The difference in An % may therefore only be due to this circumstance. As a comparison it may be mentioned that in a rather large occurrence of saussuritized anorthosite by the road Ø. Tveite the plagioclase shows 42 % An, $\alpha_{Na} \sim 1.550 \perp \gamma$. This fairly good accordance of the figures indicates that the difference in An % is due to the same saussuritization.

In the anorthosite there are some types, whose mineral assemblages deviate somewhat from the usual one.

Quartz-anorthosite. On the Valafjell there is some which contains fairly much quartz. The quartz, which is abundant, has entered later. Quartz and plagioclase are the main minerals. The dark minerals are estimated to constitute less than 12 %. The plagioclase is saussuritized; it is an oligoclase, $\alpha_{Na} \sim 1.542 \perp \gamma$, ɔ: 25 % An. It has been partly antiperthitically developed. During the saussuritization muscovite and epidote have been formed. Another dark mineral is chlorite with segregations of ore. The chlorite has lavender-blue interference colours, and is a penninite. A little brown biotite is found together with the chlorite. Some zircon occurs in idiomorphic crystals.

The rock may be regarded as a link of transition to silexite.

Of other deviating types an antiperthite rock from the Larsfonnfjell will be mentioned. Macroscopically it is impossible to distinguish it from a regular anorthosite. The feldspar is fresh, and has the characteristic tinge of light lilac. Under the microscope it displays common mortar texture with large grains in a brecciated ground mass.

The chief mineral is a plagioclase which is mainly antiperthitically developed. It has small fine spindles in regular arrangement. The grains show undulating extinction. A little common plagioclase is also found. The An % is surprisingly low. Index of refraction in section $\perp \gamma$: $1.543 > \alpha_{Na} >> 1.540$, ɔ: 24 % An.

The dark minerals are estimated to constitute 10—15 vol.%. It is noteworthy that the rock contains orthorhombic pyroxene; as mentioned above only monoclinic pyroxene has otherwise been found within the anorthosite area. The pyroxene is seldom quite fresh; as a rule there is a core of pyroxene surrounded by a marginal zone of bluish green hornblende and biotite. The freshest cores show pleochroism in pink and green: $\gamma =$ light green, $\alpha =$ pink. Beside the bluish green hornblende which has clearly been formed at the expense of the pyroxene, there is a brownish green hornblende of the same type as is otherwise common in the anorthosite area. It occurs in independent grains, and is considered primary.

Brown biotite is found in connection with the bluish green hornblende, and besides in rosettes around the ore which is an ilmenite. Some hematite occurs.

Zircon occurs in fairly large grains with fine crystal faces. One of the largest grains has been measured: 0.34 mm × 0.18 mm. As zircon does not otherwise occur in the anorthosite area, the birefringence was measured by means of Berek's compensator. First the thin section thickness was measured by means of a section $\perp \gamma$ in the plagioclase; it was 36 μ. The birefringence was determined as 0.044. The section was a somewhat oblique normal one. The birefringence is given as 0.045—0.058 according to Winchell.

This rock is distinguished from the others by the low An %, and by having orthorhombic pyroxene. It is worth noting that orthorhombic pyroxene has only been found once within the anorthosite area. In the anorthosite of the Bergen district the monoclinic pyroxene prevails over the orthorhombic one (Kolderup 1940). As to the Sogn—Jotunheimen district Goldschmidt (1916) states that the opposite is the case. The same is the case with the anorthosite of Sørlandet (Kolderup 1896).

Intermediate Rocks.

Diorite. As mentioned above these rocks are first of all found as links of transition between the anorthosite and the mangerites in the boundary regions of the anorthosite. All transitional links are here represented, from rocks approximating to gabbro to rocks approaching jotunite and mangerite. A general rule is that the plagioclase becomes more acid if the rock contains potash feldspar. Apart from the potash feldspar which is a microperthite, the mineral assemblage does not differ from that described from the boundary areas of the anorthosite. The dark mineral is as a rule a bluish green hornblende. Pyroxene also occurs, but it is seldom fresh. Most frequently it has been more or less altered to hornblende. The An % in the plagioclase is seen from the table.

No.	Locality	Index of refraction in Na-light	An %	Rock name
1	Gjeita fjell	$\alpha \sim 1.550 \perp \gamma$	41	Diorite
2	Ø Tveite	$\alpha \sim 1.550 \perp \gamma$	41	»
3	Gryte elven	$\alpha \sim 1.543 \perp \gamma$	28	»
4	Gryte elven	$\alpha \lesssim 1.543 \perp \gamma$	26	»
5	Larsfonnfjell	$\alpha \sim 1.542 \perp \gamma$	25	»
6	Brekkenipo	$\alpha \lesssim 1.542 \perp \gamma$	22	»

The last two samples have microperthite.

A volumetrical analysis has been made of No. 2:

Plagioclase	51.10	vol.%
Hornblende	33.10	»
Biotite	12.70	»
Epidote-mineral	1.77	»
Ore	1.12	»
Apatite	0.21	»

% dark minerals: 48.90.

An % in the plagioclase: 41.

The rock accordingly belongs to the diorite family.

The hornblende is not here the common bluish green one. It is purely green with pleochroism: $\gamma =$ olive green, $\beta =$ green, $\alpha =$ yellowish green. Refraction is high: $\gamma_{Na} \leqq 1.691$, $\alpha_{Na} \leqq 1.673$. According to Kunitz' tables this corresponds to a basaltic syntagmatite-hornblende with low Ti content and 35 % Fe compound. The same hornblende is found in jotunite and in more acid rocks.

Diorites occur also outside the anorthosite area. From the border area of the gabbro of Uppheimsdalen farther west two samples have been examined. They are distinguished from the gabbro by containing common bluish green hornblende. The refraction is somewhat higher than that of the brownish grey hornblende of the gabbro. $\alpha_{Na} \leqq 1.660$. c/γ 18°. They have besides a more acid plagioclase. $\alpha_{Na} \leqq 1.543 \perp \gamma$. ɔ: 25 % An for one of the two samples, and $\alpha_{Na} \geqq 1.543 \perp \gamma$, ɔ: 30 % An for the other.

By the river of Skjulsetdalen a little farther to the south of Hafsåsen there is some diorite. It has the following mineral composition:

Plagioclase	61.90	vol.%	
Biotite	27.44	»	% dark minerals: 38.10.
Uralite	9.80	»	An % in the plag.: 28 (± ½)
Clino-zoisite	0.30	»	$\alpha_{Na} \geqq 1.543$.
Apatite	0.27	»	
Garnet	0.29	»	

The uralite has the crystal form of the pyroxene and consists of an aggregate of fine hornblende needles together with small grains of clino-zoisite, which again is usually surrounded by a coating of brown biotite. In the integration it has not always been possible to distinguish the clino-zoisite from the hornblende, so that the value of the clino-zoisite becomes a little too low.

Jotunite is found in small occurrence by the farm road to Tveite. It is a massive, medium-grained rock with brown weathered surface. The feldspar has a bluish tinge.

Volumetrical analyses of the two samples give the following figures:

	I	II	
Plagioclase	36.08	54.20	*No. I.* % dark minerals: 26.32.
Microperthite	32.50	15.60	% An. in the plagioclase: 28.
Quartz	5.10	3.61	% plagioclase of total feldspar: 52.2.
Hornblende	18.20	8.30	

	I	II
Orthorhombic pyroxene	1.22	5.08
Monoclinic pyroxene	4.70	3.49
Biotite	1.00	6.16
Ore	0.70	2.67
Apatite	0.40	0.89
Titanite	0.10	—

No. II. % dark mineral: 26.59. % An. in the plagioclase: 27.5. % plagioclase of total feldspar: 77.5.

The samples are taken from the same locality, and they show that the relation between potash feldspar and plagioclase may vary greatly within a small area.

The mineral assemblage is the same as that of the mangerite, and the individual minerals will be described in detail under this section. The plagioclase is a little more basic. Measurements of the index of refraction show: I. $\alpha_{Na} \geqq 1.543$, ɔ: 28 % An ($\pm \frac{1}{2}$). II. $\alpha_{Na} \sim 1.543$, ɔ: 27.5 % An.

The orthorhombic pyroxene shows pleochroism in faintly green and pink. Index of refraction: $\alpha_{Na} \sim 1.692$. According to Winchell (1933) this corresponds to a bronzite with c. 25 % Fe. The monoclinic pyroxene is a diallage. It is light green and cleaves well after (100). The hornblende is olive green with pleochroism in yellowish green. It appears to be primary. $\alpha_{Na} \sim 1.670$, $\gamma_{Na} \sim 1.691$. According to Kunitz (1930) this corresponds to a basaltic syntagmatite-hornblende with small titanium content and c. 35 % Fe-compound (Fig. 13).

Mangerite is, as mentioned above, the predominant rock. The distribution is shown on the map. It is impossible to fix any sharp border between the igneous rocks and the gneisses. The ground is also much covered here. Usually small crags project, the rock of them being unaltered with eugranitic structure. In between these crags there are foliated zones. The strike direction is the common Caledonian of the area. The fresh mangerites may have flow lines, whose direction is NW—SE, i. e. the same direction as that of the flow lines of the gneisses. In appearance it varies a great deal. Most frequently it is brownish on weathered surfaces, and the feldspar has usually a bluish tinge. But there are also lighter types where the feldspar is quite white. It is as a rule medium-grained, seldom coarse-grained. In the massive types the dark minerals are usually assembled in lumps. Where there are flow lines they are marked by the dark minerals.

Distinction is made between pyroxene-, uralite-, amphibole-, and biotite-mangerites. Among these the uralite-mangerites are most common. Volumetrical analyses of the three typical mangerites have been perfomed. As a comparison analysis of a mangerite from Kålsås, Manger is cited. This analysis has been carried out by Asbjørn Skordal.

	I	II	III	IV
Microperthite	66.06	74.10	50.90	57.50
Microcline	—	—	11.10	—
Plagioclase	—	2.50	22.00	7.65
Quartz	4.69	—	—	1.35
Pyroxene	23.70	—	—	5.71
Hornblende	—	16.70	14.40	2.98
Biotite	1.34	5.50	—	16.20
Ore	3.89	0.30	1.25	5.36
Apatite	0.10	0.20	0.35	0.96
Zircon	0.22	—	—	0.14
Titanite	—	0.70	—	0.66
Epidote-mineral	—	—	—	0.93

1. Pyroxene-mangerite from Tveite, Vossestrand.
2. Amphibole-mangerite from Ettesvoll, Vossestrand.
3. Uralite-mangerite from Valafjell, Vossestrand.
4. Pyroxene-mangerite from Kålsås, Manger.

% dark minerals: 29.25 — 23.40 — 16.0 — 33.50 respectively.

Characteristic of the mangerite is that the potash feldspar prevails over the plagioclase. The analyses show that the relation may vary quite a lot. As to % plagioclase of total feldspar we get the following figures for the four samples: 0.0 — 3.2 — 26.1 — 11.8.

In the typical mangerites the potash feldspar is always microperthitic. The size and form of the blebs vary somewhat, but lengthened, slender spindles tapering at the ends are most common, and they are quite regular. The following sizes are common:

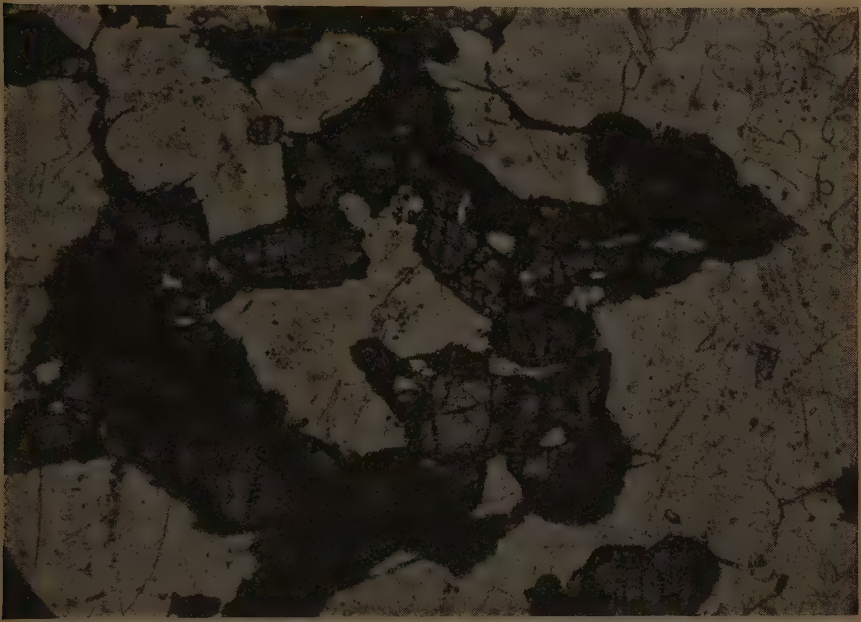


Fig. 13. Jotunite, Tveite with ortho-rhombic pyroxene and ilmenite with coronas of biotite. × 35.

0.025 mm × 0.175 mm
0.0125 » × 0.100 »
0.0100 » × 0.500 »

The latter is the most common.

When the spindles are so large, it is possible to measure their An % by means of immersion liquids. For the four analyzed samples we get the following values:

No.	Index of refraction in Na-light	An.% in the plagioclase	Index of refraction in Na-light	An.% in the spindles	Locality
1			$\alpha \sim 1.540$	22	Tveite
2	$\alpha \leqq 1.540$	21	$\alpha \leqq 1.540$	21	Ettevoll
3	$\alpha \sim 1.563$	16	$\alpha \geqq 1.531$	10	Valafjell
4		26		26	Kålsås

In No. 3 the plagioclase is somewhat altered, and the spindles are a little more altered than the plagioclase outside. The mangerite from

Manger has a somewhat more basic plagioclase; its An % has been examined by means of a universal stage by Mr. Skordal. The plagioclase from pyroxene-mangerites from the Bergen district contains about 20 % An (Kolderup 1940).

Here the data for some other samples will be given:

No.	Index of refraction in Na-light	An.% in the plagioclase	Index of refraction in Na-light	An.% in the spindles	Locality
5	$\alpha \sim 1.540$	22	$\alpha \sim 1.540$	22	Brekknipo
6	$\alpha \geq 1.535$	16	$\alpha \geq 1.532$	13	Kjednarhovden
7	$\alpha \leq 1.540$	21			Tveite
8	$\alpha \geq 1.543$	14.5			Rygg
9	$\alpha \sim 1.536$	16.5			Løen, Østre

As will be seen the An % of the spindles and of the plagioclase outside may vary somewhat, first of all when the plagioclase is somewhat altered. It may then be due to different degrees of saussuritization in and without the spindles. The maximum uncertainty of the measurements is ± 1 %, so that this cannot explain the difference. It may be stated that H. J. Alling (1938) in investigations of perthites has found out that the An % within each spindle may vary by up to 4 %.

As mentioned above the spindles are lengthened and slender, and though being much longer, they mostly resemble what Alling (1938) calls "string perthites". These long spindles lie in dense arrangement, and constitute a very great part of the volume of the grain. In order to illustrate this numerically some microperthite grains have been volumetrically analyzed.

The first two grains are from sample No. 2, and the third is from No. 1. The analysis gave the following result:

No.	% Potash feldspar	% Plagioclase	Locality
1	49.5	50.5	Ettesvoll (2)
2	42.5	57.5	» »
3	41.0	59.0	Tveite (1)

The measurements were performed in sections as nearly as possible vertical to the diametral plane of the spindles, and the integration was undertaken vertically to the longest axis. Such a measurement is not quite exact, because it is very difficult to adjust the screw on the integration table exactly when such small distances are concerned. But the measurement gives at least an idea of the relative quantities of the two feldspars.

In the systematic placing of the mangerite microperthite was considered as potash feldspar. If in the four examined samples we consider the microperthite as 50 % plagioclase and 50 % K-feldspar, the result is that we get plagioclase as the predominating feldspar in two of the samples. That means they would fall within the syenodiorite family instead of the syenite family. Theoretically it would prove most satisfactory to take this into account in a systematic placing, but it can hardly be carried out in practice. It is only very few microperthite grains that are suited for volume analysis, and the form and size of the spindles vary so much that grains from a series of samples would have to be volumetrically analysed in order to obtain a satisfactory result. But even though it cannot be taken into account in the systematization, it is all the same valuable to be aware of the fact.

Besides microperthite, microcline and microcline-microperthite occur in some mangerites. They are not found in quite fresh types, and are regarded as altered products of the microperthite.

Most frequently they also contain some quartz.

As mentioned above, uralite (Fig. 14) is the most common dark mineral, but pyroxene also occurs.

Orthorhombic pyroxene may be found in the mangerites as cores in the uralite grains. Quite fresh it occurs in the jotunite from Tveite. Characteristic is a distinct pleochroism in pink and green, parallel extinction on prisms and pinacoid faces, and distinct cleavages forming angles of c. 90° with each other. Index of refraction: $\alpha_{Na} \sim 1.692$. According to Winchell this corresponds to a hyperstene with c. 25 % $FeSiO_3$, ɔ: a bronzite.

As a rule it has a marginal zone of bluish green hornblende, often mixed with clino-zoisite. Biotite, too, may occur here. There are all the transitions from such a fresh pyroxene with a narrow marginal zone of hornblende to typical uralites, where only the crystal form of the pyroxene has been preserved. It looks as if the orthorhombic pyroxene is first of all subject to uralitization, but most frequently the transformation has proceeded so far that it is impossible to decide whether the pyroxene has been orthorhombic or monoclinic. In gabbro at Ettesvoll there was uralitization of monocline pyroxene.

Monoclinic pyroxene is not so common among the mangerites. Where it occurs it is light green of colour and has a pronounced cleavage after (100). c/γ c. 42°. $\gamma_{Na} \sim 1.716$. This corresponds with a diallage.

As mentioned above hornblende occurs as a marginal zone around the pyroxene and in the uralites. This hornblende is common bluish green with strong pleochrism: $\alpha =$ yellow, $\beta =$ green, and $\gamma =$ bluish green. Most frequently it occurs in intimate mixture with clino-zoisite, often also with biotite, and is difficult to determine. In a mangerite-syenite from Ulvund the hornblende grains are relatively large and pure. The following determinations were made: c/γ 15°, $\alpha_{Na} \sim 1.653$. According to Kunitz (1930) this corresponds to a syntagmatite-hornblende with 36 % Fe compound. Only this secondary hornblende occurs in the mangerites. It is sharply distinguished from the purely green hornblende of the jotunite which occurs in large grains together with pyroxene, and is regarded as primary.

The biotite may be both brown and green; the colour may change in the same thin section. Characteristic is its occurrence in rosettes, either alone or as a marginal formation around ore. These biotite rosettes are typical of the Bergen—Jotunheimen tribe; they have been described both from the Bergen district and from Jotunheimen. The biotite may also occur in connection with hornblende and an epidote-mineral as a marginal zone around the pyroxene. The mangerite from Brekkenipo contains much biotite. It is brown with strong pleochroism: $\alpha =$ light yellow, $\beta = \gamma =$ dark brown. It occurs in large independent flakes, and besides as fine small needles in the marginal zone around the orthorhombic pyroxene.

The ore is, as a rule, an ilmenite, and often encircled by a margin of titanite. Otherwise rosettes of biotite are, as mentioned above, common. Magnetite also occurs. In accessory amounts are found: Apatite, titanite, and zircon. An epidote-mineral is also common.

It is worth noting that the mangerites and also all the unaltered igneous rocks are primarily found in the immediate vicinity of the

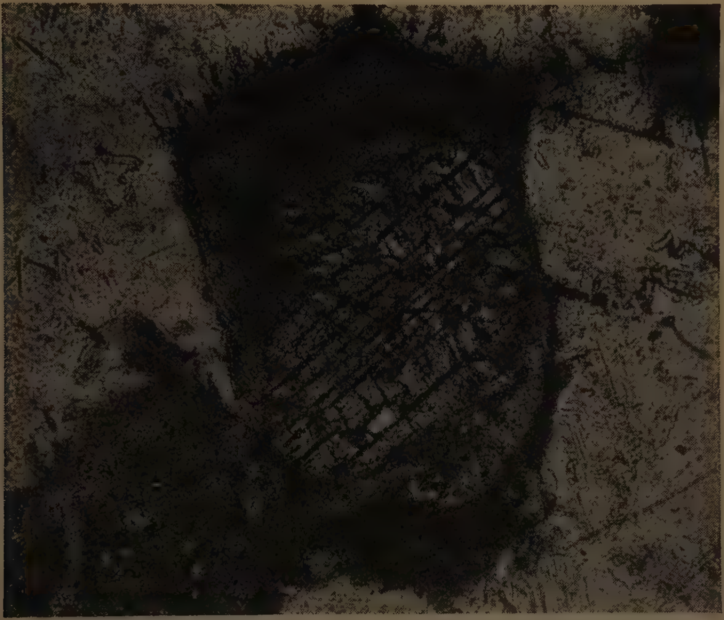


Fig. 14. Uralite from mangerite SE Valafjell. × 140.

anorthosite. Especially between the two anorthosite massifs southeast of Valafjell the rocks have been well preserved.

The sole exception from this rule is a small area of mangerite and *mangerite-syenite* at Ulvund at Myrkdalsvannet. It lies immediately above the phyllite. In the border region itself the rock is highly brecciated. There are mylonites and mylonite-gneisses, but already a couple of meters from the boundary the rock is comparatively unaltered. Petrographically this rock is here transitional between mangerite and mangerite-syenite. It is much lighter than the common mangerite, the colour being greyish to white. The rock is medium- to coarse-grained, and contains few dark minerals, which are assembled in lumps.

A volume analysis of a sample from this area has been performed:

Microperthite	75.00	vol.%	
Plagioclase	11.30	»	% dark minerals: 1.16.
Quartz	12.20	»	% plag. of total feldspar: 13.
Calcite	0.34	»	% An in the plag.: 9.
Hornblende	1.16	»	

The integration here gives a false picture of the relation between the light and the dark minerals. The rock is comparatively coarse-grained, and the dark minerals are assembled in lumps, and it is not easy to get a complete representation of them in a thin section. But on the average the dark minerals certainly constitute more than 15 %. There is very much quartz, but part of it appears to have been newly formed.

The An % in the plagioclase and in the spindles of the microperthite varies from 10—15 %. The following measurements have been made:

$\alpha_{Na} \leqq 1{,}535$, ɔ: 15 % An. in the spindles $\alpha_{Na} \geqq 1{,}532$, ɔ: 11 % An. Ulvund.
$\alpha_{Na} < 1{,}529$, ɔ: 8 % „ —„— $\alpha_{Na} \sim 1{,}532$, ɔ: 10 % „ „
$\alpha_{Na} \sim 1{,}534$, ɔ: 14 % „ „

The plagioclase is more acid than in the typical mangerites, and the majority of the samples are nearer to mangerite-syenite.

The dark mineral is usually a bluish green syntagmatite-hornblende: $\alpha_{Na} = 1.653$ ($\pm$ 0.001). In some thin sections it appears clearly that it has been formed at the expense of the pyroxene. It is also altered to brown biotite.

Pyroxene occurs; in a sample from Ulvund a diallage with $\alpha_{Na} \sim 1.694$ has been determined.

Some calcite has been formed secondarily.

Otherwise the mineral assemblage is the same as that of the mangerite.

Acid Rocks.

Charnokite. On the road to Haugstøl east of Uppheimsvatn a little area of charnockite has been found. It is a grey, medium-grained rock.

A volume analysis shows the following values:

Microperthite	35.24	vol.%	
Plagioclase	25.20	»	
Quartz	27.80	»	% dark minerals: 11.8.
Pyroxene, orthorhombic	4.50	»	% plag. of total feldspar: 42.
Hornblende	2.34	»	% An in the plag.: 25.
Biotite	2.60	»	
Ore, (ilmenite)	1.76	»	
Apatite	0.56	»	

The rock contains too much plagioclase for a typical charnockite, but it falls decidedly within the frame of the charnockite family as this is defined.

The plagioclase is an oligoclase, $\alpha_{Na} \sim 1.542$, ɔ: 25 % An.

The hornblende is olive green, and the same type as that described from the jotunite of Tveite: a syntagmatite-hornblende with small titanium content and c. 36 = Fe-compound.

The pyroxene is a bronzite, $\gamma_{Na} \sim 1.704$, $\alpha_{Na} \sim 1.691$. According to winchell this corresponds to c. 25 % $FeSiO_3$.

The biotite is brown, and is found in rosettes around the ilmenite.

The microperthite is the regular spindle perthite.

Between the quartz and the potash feldspar, a little myrmecite has been formed. Apatite is found accessorily in small rounded-off grains.

Birkremite is found in several localities. It is more common than the charnockite. Distinction is made between amphibole- and biotite-birkremite.

As a type of the former a reddish, medium-grained rock from the south slope of the Valafjell will be described:

A volume analysis shows the following figures:

Microperthite	48.20	vol.%
Plagioclase	19.40	»
Quartz	27.49	»
Hornblende	0.87	»
Biotite	0.36	»
Ore. (ilmenite)	3.05	»
Epidote-mineral	0.36	»
Titanite and zircon ...	0.27	»

% dark minerals: 4.91.
% plag. of total feldspar: 29.0.
% An in the plag.: 8.

The plagioclase is a basic albite, $\alpha_{Na} \sim 1.530$, ɔ: 8 % An.

The plagioclase of the microperthite spindles: $\alpha_{Na} \geq 1.530$, ɔ: 8—9 % An.

The microperthite has the common, long spindles.

The hornblende is bluish green, partly altered to chlorite.

The biotite is light green, and occurs in characteristic rosettes, often around the ilmenite.

The epidote-mineral is an orthite-epidote. V. M. Goldschmidt (1916) mentions that it is a mineral characteristic of granites of the Bergen—Jotunheimen tribe.

The biotite-birkremites do not contain hornblende, but otherwise the mineral assemblage is the same. The birkremites have most usually a reddish colour, and are thereby distinguished from the mangerites which have a more brownish tinge.

Enderbite. In the hillsides east of the anorthosite massif at Kamben a little enderbite has been found. The rock has the Caledonian schistosity, but the minerals are relatively well preserved.

Plagioclase is the only feldspar, it is partly antiperthitically developed. It is an oligoclase, $\alpha_{Na} \sim 1.539$ ($\pm$ 0.002), ɔ: c. 21 % An. It has undulating extinction. Large grains are found in a comparatively fine ground mass which consists of quartz and feldspar. The ground mass is so fine-grained that it is not possible to get a satisfactory integration of the rock constituents.

Quartz is estimated to constitue more than 12 %. Further light green pyroxene occurs. It is monoclinic, cleaves well after (100), but is highly crushed. It has been partly altered to bluish green hornblende and chlorite. In accessory amounts are found: ore and titanite. Epidote has been formed secondarily.

Silexite. As mentioned above zones and dikes rich in quartz are very common, especially in the anorthosite area. Within this again they are mostly found in the banded rocks in Hodnardalen. Otherwise silexite is very common in the boundary regions between anorthosite and mangerite. It is also found outside the anorthosite area in the mangeritic rocks there. The zones may have a considerable extension. On the western slope of Gjeitafjell there is a zone that is about 10 m thick and about 300 m long. The boundary between the silexite and the adjacent rock is sharp. In several places it has been observed that silexite borders on eclogite or pyroxenite, these being then biotized close to the boundary. Similar zones rich in quartz have been described from the Bergen district (Kolderup 1940), but their origin is an open question. It has, however, also been mentioned that they may be remnants of sediments. With this possibility in mind I also searched here for traces of bedding, strati-

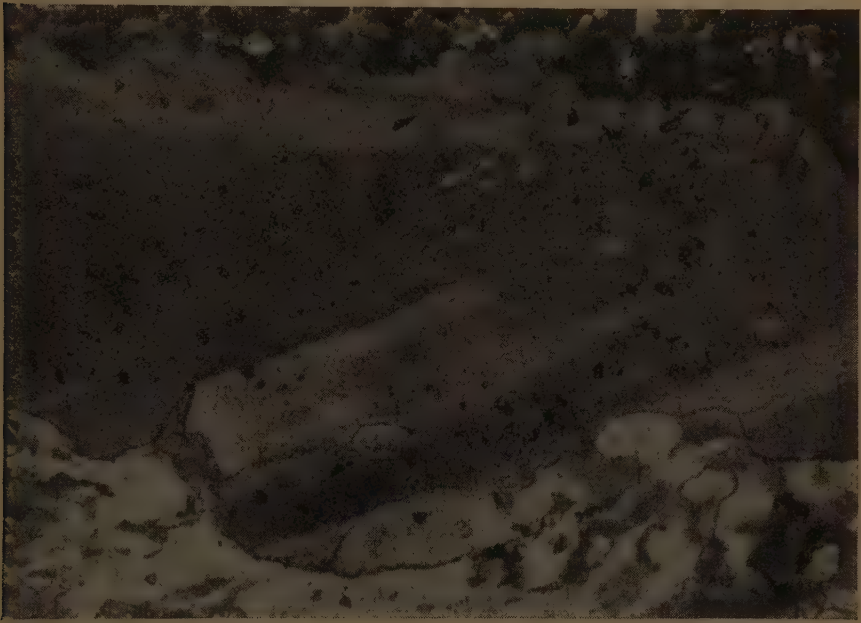


Fig. 15. Microperthite in silexite Tveite. × 120.

fication, etc., but with no result. On the contrary, during the microscopic examination of the collected material I was convinced that the rock in question is aqueo-igneous or hydrothermal quartz, that is silexite.

Observations in the field also support this assumption. Of such observations the following will be mentioned: Biotitization around the eclogites; the fact that the silexite occurs so extensively in the border zone between anorthosite and mangerite. It is reasonable that aqueo-igneous quartz dikes exhibit a preference for border zones which are, as a rule, weak zones, but there is no reason why possible remnants of sediments should exactly be found along the boundary. Here may be mentioned that a description from the Bergen district (Kolderup 1941) shows that also here the zones rich in quartz occur in the vicinity of the borders of the anorthosite. In the banded rock of Hodnardalen the silexite bands are parallel to the other bands, and when it occurs in the anorthosite, the dikes run parallel to its foliation. These are all facts in favour of a hydrothermal or aqueo-igneous origin.

Microscopical examinations. Four thin sections of silexite from different parts of the anorthosite area were examined. Common to all is that the quantity of quartz amounts to some 90 %. Besides, all contain some feldspar, three of the samples have both plagioclase and potash feldspar, the fourth only plagioclase. The potash feldspar is especially interesting: it is in all the cases a microperthite. The spindles are very small, regular, and have clearly higher indices of refraction than the ground mass. Their length averages 0.025 mm, i. e. somewhat larger than in what Alling (1938) has called "stringlet perthites". These, according to Alling, have been formed by exsolution. The fact that microperthite grains are found in the silexite is also in favour of a "magmatic" origin. The grain in the figure is albitized at the edges, and in the interior of the grain there are some comparatively large irregular spindles that must have been formed later, by replacement. Similar microperthite grains are found in the two other samples (Fig. 15).

A sample from the cross-roads at Tveite has been volumetrically analyzed. It shows the following mineral composition:

Quartz	91.90	vol.%	
Plagioclase	6.80	»	
Microperthite	0.32	»	% dark minerals: 0.98.
Biotite	0.16	»	% plag. of total feldspar: 96.1.
Muscovite	0.31	»	
Epidote	0.32	»	
Titanite	0.19	»	

The rock is consequently a typical silexite. The sample from Tveite is strongly pressed. Under the microscope a pronounced orientation of the grains is seen.

The most interesting fact about this sample, apart from the microperthite, is the titanite. Some few large grains of titanite occur, and they exhibit distinct crystal form. The largest grain has been measured, and is 0.36 × 0.9 mm. That the grains have preserved their crystal form is also greatly in favour of an aqueo-igneous or hydrothermal origin.

Other minerals occurring in the silexite are: Hornblende, epidote, orthite-epidote, garnet, muscovite, chlorite, and ore. That is the same mineral assemblage oc-

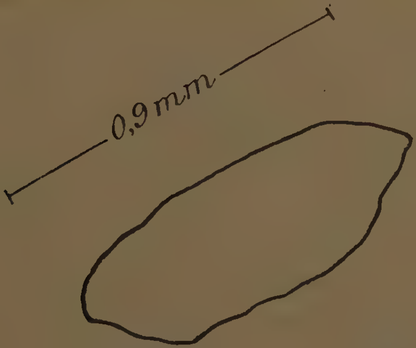


Fig. 16. Micro-drawing of titanite in silexite.

curring in the other intermediate and acid rocks. Also this is indicative of a connection between them and makes for considering the rocks rich in quartz as the last acid link of the eruption.

Pegmatite.

It has been mentioned above that pegmatite dikes cutting the flow lines transversally have been found on Valafjell. Pegmatite dikes are otherwise very common, mostly in the anorthosite areas, but they are also found outside in the mangeritic rocks. Two generations may be distinguished (see p. 149), one before and one after the foldings in the anorthosite.

The oldest are most common. The main minerals are: Quartz, plagioclase, and muscovite. The plagioclase is generally an oligoclase. In a sample from the Valafjell the index of refraction has been measured: $\alpha_{Na} \sim 1.541$. ɔ: 23—24 % An. It may also be somewhat more basic. By the road cutting at Lunde has has been measured: $\alpha_{Na} \leqq 1.550$. ɔ: c. 40 % An.

Other minerals are: a mineral of the clino-zoisite—epidote series, apatite, garnet, and titanite.

The pegmatites are not particularly coarse-grained. The muscovite flakes seldom become larger than 2—3 cm. The largest titanite grain is not even as large as a millimetre; a measurement shows 0.72 × 0.118 mm.

The youngest pegmatites are distinguished from the others by containing potash feldspar. There is by a cross-road at Tveite such a pegmatite dike which clearly has been formed later than the foldings.

The mineral composition is otherwise the same, and the plagioclase is an oligoclase.

In the mangerite area north of Uppheimsvannet there has also been found pegmatite with potash feldspar. Here are no foldings.

Diabase.

By the cross-road at Tveite a diabase dike cutting through silexite has been found. It is dark and fine-grained. The main minerals are hornblende and plagioclase. The plagioclase has zonal texture. The hornblende is a common bluish green one. In subordinate amounts are found: Epidote and quartz. The border towards the silexite is sharp.

Facies of the Rocks of the Anorthosite Kindred.

The following mineral assemblage is characteristic of the interior of the anorthosite massif: Labradorite, eclogite-garnet, diallage, and brownish green hornblende. This mainly corresponds to the granulite facies, but there is no complete accordance in the mineral assemblage. Hyperstene is lacking, and so is rutile, which is the characteristic titanium mineral of the granulite-facies. Ilmenite is the only titanium mineral; titanite has not been found here. But the minerals also exhibit many features characteristic of the granulite-facies. The diallage is often light green — it may also be colourless. The hornblende is brownish or greyish green. The plagioclase may be antiperthitically developed, but this is not common. Most characteristic is the garnet rich in pyrope that belongs to the eclogite- and granulite-facies only. The mineral assemblage, at any rate, is nearer to the granulite-facies than to the hornblende-gabbro-facies.

In the boundary regions of the anorthosite the conditions are somewhat different. The An % in the plagioclase decreases towards the boundary, and at the same time there is an increase of the quantity of epidote. The pyroxene disappears, and instead a bluish green hornblende occurs, which is plainly secondary. The eclogite-garnet is also found in the boundary regions, but much indicates that it is no longer stable. The garnet occurs usually together with epidote, biotite, and hornblende. In the interior these minerals may be lacking, or only consitute a narrow zone around the garnet. In the boundary regions, however, these minerals constantly gain ground at the expense of the garnet, so that, finally, only narrow streaks are left. The critical minerals here are therefore: Plagioclase (andesine), hornblende, and remnants of pyrope-garnet. If this is considered an unstable relic, the rocks of the boundary region belong to the hornblende-gabbro-facies. In all cases, they are nearer to the hornblende-gabbro-facies than to the granulite-facies. Both the hornblende and the more acid plagioclase show that the temperature must have been lower in the boundary zones than in the interior. *The rocks outside the anorthosite area* are, as mentioned above, chiefly intermediate and acid rocks. The essential difference of the mineral assemblage is that the pyrope-garnet is lacking here. Orthorhombic pyroxene occurs in the freshest types. It has often pleochroism in green and pink, as is characteristic of the granulite facies. Diallage occurs, but relatively seldom, and its colour is light green. But otherwise uralites of every stage of development and bluish green hornblende constitute the majority of the dark minerals. Of titanium minerals both ilmenite and titanite are nearly always present. The plagioclase is, as a rule, an oligoclase, and the potash feldspar is microperthitically developed.

This is a mineral assemblage which has features in common both with the granulite-facies and the hornblende-gabbro-facies. Types that fit into the granulite-facies are found, but these are numerically few. Fresh pyroxene — both orthorhombic and monoclinic — is rare, and the majority of these rocks will fall within the hornblende-gabbro-facies.

Chemical analyses of the rocks have not been made for this work. As it was desirable to place the various rock types in an ACF diagram, the chemical composition and the ACF values have been calculated on the basis of the volumetric analyses. The reliability of such calculations have been discussed. Shand (1927) quotes an analysis and a calculation of a granite, the accordance is very good. He says: "It shows that it is possible to deduce the chemical composition of a rock from its mineralogical composition more accurately than one can perform the reverse operation of deducing the mineralogical composition from a chemical analysis of the rock."

No chemical analyses are available of the rocks that have been volumetrically analysed and calculated in this work, but the calculated values have been compared with analyses of corresponding rocks from the Bergen and Jotunheimen districts. They exhibit a good accordance.

In the re-calculation von Philipsborn's tables (1933) have been used. In the calculation of hornblendes and pyroxenes analyses of minerals from corresponding rocks have been made use of, when such have been available.

Analyses and corresponding ACF values have been calculated for four samples, and besides for two garnets.

Before the re-calculation to molecular quotients the calculations have been corrected for ilmenite and apatite. This has also been done for biotite; the figures A' C' F' indicate the values after this correction.

No. 1. *Pyroxene-mangerite, Tveite.*

	Vol. %	G. %	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	P_2O_5	ZrO_2
Plagioclase	39.48	36.32	22.80		8.51				1.70	3.31				
K-feldspar	26.58	23.45	15.15		4.32						3.98			
Quarz	4.69	4.37	4.37											
Diallage	23.73	27.86	13.40	0.20	2.10	2.20	1.50	3.50	4.46	0.40	0.10			
Biotite	1.31	1.39	0.56			0.26		0.38			0.14	0.05		
Ilmenite	3.89	6.12		3.21			2.91							
Zircon	0.22	0.37	0.12											0.25
Apatite	0.10	0.12							0.06				0.06	
Sum:	100.00	100.00	56.40	3.41	14.93	2.46	4.41	3.88	6.22	3.71	4.22	0.05	0.06	0.25
Molquot. × 1000:					146.50	15.40	20.90	96.20	109.80	59.90	44.80			

A = 20.18 C = 38.60 F = 41.22

A' = 20.80 C' = 40.00 F' = 39.20

(volume analysis and description v. p. 172).

Microperthite is here the only feldspar. It has large spindles of plagioclase. By the integration of the microperthite has been found 59 % plagioclase with 22 % An and 41 % potash feldspar. These values have been used in the calculation. The diallage has been calculated on the basis of an analysis of a diallage from the Bergen field (N. H. Kolderup 1925).

No. 2. *Quartz-jotunite, Tveite.*

The feldspar is here both plagioclase and microperthite. The microperthite contains a few drop-like plagioclase spindles. The content of plagioclase in the potash feldspar has been estimated at 5 %. The plagioclase contains 25 % An.

The diallage has been calculated according to the same analysis as the diallage in No. 1. The hornblende has been calculated on the basis of an analysis by Kunitz (1930) of a hornblende with the same index of refraction. The orthorhombic pyroxene has been determined as a bronzite with 25 % Fe SiO_3. The determination has been made according to Winchell's tables on the basis of the index of refraction.

	Vol. %	G. %	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	H_2O	P_2O_5
Plagioclase	42.58	40.55	24.85		14.39					2 35	3.40			
K-feldspar	26.00	23.65	15.20									4.01		
Quarz	5.10	4.85	4.85											
Diallage	4.50	5.54	2.66	0.05	0.42	0.45	0 30		0.71	0.85	0.08	0.02		
Broncite	1.22	1.44	0.85				0.20		0.39					
Hornblende	18.40	21.20	8.60	0.61	1.88	0,42	1.07	0.01	3.96	2.55	0.47	0.21	1.42	
Biotite	1.00	1.06	0.46			0.18			0.28			0.11	0.03	
Ilmenite	0.70	1.13		0.60			0.53							
Apatite	0.40	0 47								0.24				0.23
Titanite	0.10	0.11	0.04	0.05						0 02				
Sum:	100.00	100.00	57.51	1.31	16.69	1.05	2.10	0.01	5.34	6.01	3.95	4.35	1.45	0.23
Molquot. × 1000:					163.70	6.60	21.90	0.01	13.24	102.50	63.70	46.20		

A = 19.1 C = 32.4 F = 48.5

A′ = 19.4 C′ = 32.9 F′ = 47.7

(vol. anal. and description v. p. 170).

No. 3. *Charnockite, Haugstøl.*

	Vol. %	G. %	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	P_2O_5
Plagioclase	27.02	26.60	16.50		6.50				1.40	2.20			
K-feldspar	33.46	31.40	20.40		5.70						5.30		
Quartz	27.80	27.60	27.60										
Broncite	4.50	5.50	3.03				0.76	1.71					
Hornblende	2.34	2.50	1.01	0.06	0.22	0.05	0.10	0.46	0.30	0.05	0.02	0.23	
Biotite	2 60	2.90	1.10			0 50		0.71			0.29	0.30	
Ilmenite	1.76	3.00		1.58			1.42						
Apatite	0.52	0.50							0.25				0.25
Sum:	100.00	100.00	69.64	1.64	12.42	0.55	2.28	2.88	1.95	2.25	5.61	0.53	0.25
Molquot. × 1000:					122.00	3.40	12.00	71.40	30.30	36.30	59.61		

A = 20.6 C = 20.9 F = 58.5

A′ = 25.2 C′ = 23.6 F′ = 51.2

(Vol. anal. and description v. p. 178).

The microperthite has in the calculation been estimated to contain 5 % plagioclase. The plagioclase has 25 % An. The orthorhombic pyroxene has been determined as a bronzite with 25 % $FeSO_3$, and the determination performed according to Winchell's tables on the basis of the index of refraction. The hornblende has been determined according to the same analysis as the hornblende in No. 2.

No. 4. *Anorthosite, Larsfonnfjell.*

	Vol. %	G. %	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	P_2O_5
Plagioclase	92 00	90.00	47.06		27.47				11.68	3.79			
Garnet	2.60	3 50	1.41		0.82		0 50	0.37	0.43				
Hornblende	0.90	1.30	0.53		0.09	0.05	0.14	0 21	0.17	0 02	0.01	0.05	
Biotite	2.80	3 30	1.34			0.62		0.88			0.30	0.13	
Epidote-min.	1.60	1.80	0.72		0.61				0.44			0.03	
Scapolite	0.10	0 10	0.05	0.01					0.01	0.01			
Sum:	100.00	100.00	51.11	0.01	29.00	0.67	0.64	1.46	12.73	3.82	0.31	0.21	
Molquot. × 1000:					284.50	4.20	8.90	36.20	227.00	61.60	3.30		

A = 45.10 C = 45.70 F = 9.20

A′ = 47.10 C′ = 48.00 F′ = 4.90

(Vol. anal. and description v. p. 166).

The plagioclase has an An % of 63. The garnet has been calculated on the basis of the composition $Fe_{31}Mg_{44}Ca_{25}$. In a sample of an anorthosite-gabbro near by (c. 100 m distant) a garnet of this composition has been determined, and the composition calculated according to Eskola's tables. The hornblende has been determined according to an analysis by Kunitz (1930) of a hornblende with the same index of refraction.

No. 5. *Garnet from anorthosite-gabbro, Larsfonnfjell.*
Composition determined by means of refractive index: $Fe_{31}Mg_{44}Ca_{25}$.

	G. %	SiO_2	Al_2O_3	FeO	MgO	CaO
Pyrope	44	19 60	11 15		13.25	
Almandine	31	11.20	6.40	13.40		
Grossular	25	10.00	5.65			9.35
Molquot. × 1000:			227.60	186.50	328.60	167.70

A = 25.0 C = 18.4 F = 56.6

No. 6. *Garnet from segregation in anorthosite, Larsfonnfjell.* Composition determined by means of refractive index: $Fe_{38}Mg_{44}Ca_{18}$.

	G.%	SiO_2	Al_2O_3	FeO	MgO	CaO
Pyrope	44	19 60	11.15		13.25	
Almandine	38	13.90	7.70	16.40		
Grossular	18	7.20	4.05			6.75
Molquot. × 1000:			224.60	228.30	328.60	120.40

A = 25.0 C = 13.3 F = 61.7

A consideration of the ACF diagram of the granulite-facies shows that all the samples calculated fit well into it (Fig. 17).

This was also to be expected according to the mineral assemblage. Eskola (1939) says: "Nach ihrem Mineralbestand können alle typischen Charnockite in der Tat zur Granulitfazies gerechnet werden. Diesen nahe stehen die Mangerite und andere pyroxenführende Erstarrungsgesteine."

As to the selection of the samples calculated some remarks must be given: The mineral assemblage in the anorthosite corresponds mainly to that which is found in the interior of the anorthosite massif, but types containing monoclinic pyroxene besides hornblende are equally common. These rocks will fall within the triangle plagioclase—diopside—garnet, as has been indicated on the diagram.

As to jotunite, mangerite, and charnockite, particularly unaltered types have been selected, because these were suited for volumetrical analysis and calculation. They are not, however, typical representatives of the intermediate and acid rocks. As mentioned above the pyroxene has been usually more or less uralitized. A volumetrical analysis of such an uralite-mangerite has been performed, but it is impossible to perform a satisfactory analysis without knowing the chemical composition of the uralite.

The conclusion must be: The rocks of the interior of the anorthosite area belong to the granulite-facies while those of the boundary areas are more related to the hornblende-gabbro facies. As to the intermediate and acid rocks some of the freshest types belong to the granulite-facies, but roughly these rocks are more related to the hornblende-gabbro-facies.

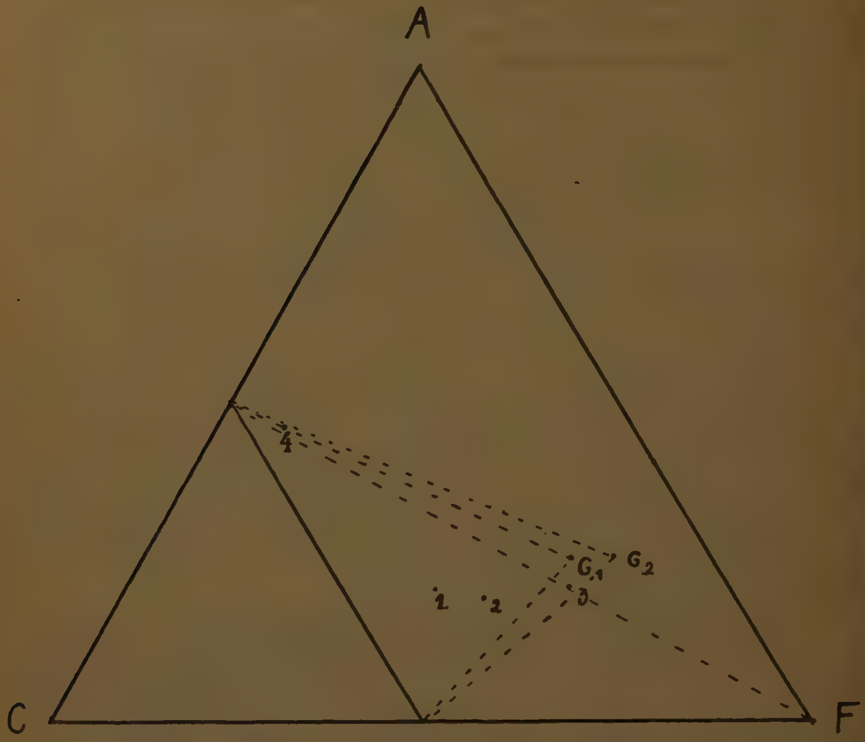


Fig. 17. Eskolas ACF diagram of the granulite facies.

1. Pyroxene-mangerite, Tveite. 2. Quartz-jotunite, Tveite.
3. Charnockite NW Haugstøl. 4. Anorthosite, Larsfonnfjell.
G_1. Garnet from anorthosite-gabbro, Larsfonnfjell.
G_2. Garnet from segregation in anorthosite, Larsfonnfjell.

We shall now see what we may conclude as to the T- and P-conditions during the formation of the rocks. In the first place it can be stated that there has been a decrease of the temperature from the interior of the anorthosite massif towards the boundary. As mentioned above the plagioclase is more acid in the boundary regions and the rock also contains more epidote. It is impossible to fix a sharp line between the boundary region and the interior of the anorthosite massif: there is a continuous transition.

A sample from the inner part of this transitional zone contains idiomorphice pidote beside plagioclase with An 63 (v. p. 167). The mineral assemblage mainly corresponds to that which is found in the interior of the anorthosite massif. This should according to Ramberg (1943) correspond to a temperature of more, than 500 degrees. In the boundary regions idiomorphic epidote is found together with plagioclase with An % 40—50, which should, according to the same author, correspond to a temperature somewhat below 500 degrees.

The eclogite-garnet shows that there must have been a high pressure in the interior of the anorthosite massif, too.

The conditions in and about the anorthosite massif correspond to those found in a regular plutone, with the highest temperatures in the interior and the lowest towards the boundary. The only thing by which it is distinguished from a common igneous rock is the excessive crushing.

We must imagine that the T- and P-conditions during the consolidation moved from the granulite-facies towards hornblende-gabbro-facies. This is especially well illustrated in the intermediate and acid rocks, because we here find a great many sharply developed reaction zones. Such late-magmatic processes indicate that the T- and P-conditions have changed during the consolidation of the rocks. It is not always an easy task, however, to distinguish between magmatic and metamorphic. The mode of occurrence of the reaction zones gives every reason to believe that they are true late-magmatic reactions.

Most common is the uralitization of the pyroxene. Both hyperstene and diallage are, as a rule, surrounded by hornblende. These zones may be very differently developed. They may consist of fine needles of hornblende standing with their longest axes vertically on the pyroxene grain, but the marginal zone consists most frequently of several hornblende grains without any definite orientation. Epidote and biotite are then often found in intimate mixture with the hornblende.

Another fact which also shows the variation of the T- and P-conditions during the consolidation are the reaction zones around the ilmenite. Ilmenite is the characteristic titanium mineral of the granulite-facies and it can be observed how the quantity of titanite increases at the expense of the ilmenite as the rock approaches the hornblende-gabbro-facies.

Biotite often occurs in rosettes around the ilmenite; these may be regarded as reaction zones.

Within the anorthosite area there are reaction zones, especially in the ultra-basic and basic rocks. Such a reaction zone of hornblende around the garnet has been described from a pyroxenite at Tveite (v. p. 160). Being, however, most frequently poorly developed it is not possible to determine the minerals they contain.

Genesis.

Hitherto we have described the igneous rocks of the anorthosite kindred without including features due to later metamorphism. As mentioned in the Introduction it is precisely these igneous rocks that are of importance to the understanding of the genesis of the rocks. We will disregard the effects of the metamorphism, and see what our description of the rocks can tell about their history.

It cannot be expected to get a clear picture of the genesis before the whole anorthosite area has been better investigated. The area described in this paper is merely a small part of the great Sogn—Jotunheimen massif, nevertheless it may relate something about the genesis of the rocks.

The prevailing opinion thus far has been that the rocks of the Sogn—Jotunheimen district have been formed by magmatic differentiation; Goldschmidt (1916). Formerly it was also believed that the rocks of the anorthosite kindred of the Egersund district had been formed in this manner, but more recently it has been discussed whether these rocks really have been formed by differentiation of a magma (Barth 1941). As to the rocks of the anorthosite kindred in the Sogn—Jotunheimen district no other theory than that of Goldschmidt has been advanced about their mode of formation. The results of these investigations support this interpretation.

We would first point to a great many features indicating that the rocks must be co-magmatic.

It must be regarded as quite certain that the intermediate and acid rocks outside the anorthosite area are genetically related. Observations in the field show that various types are intimately connected in such a manner that it is impossible to draw any line of division between them. Microscopical examinations demonstrate that the different types are connected by transition, and that broadly speaking the mineral assemblage is the same with basic microperthite as the characteristic mineral.

As mentioned above mangerite dominates here; the other types are subordinate and may be regarded merely as variations of the mangerite.

Within the mapped area no regular distribution of the various types has been found. From observations outside the mapped area I have the impression that acid rocks predominate in the region southeast of Uppheimsvatn, between the latter and the anorthosite area at Kamben. It is possible that a further investigation of this region would provide more information about the differentiation within the intermediate and acid links. The only thing that may so far be said is that the rocks must be co-magmatic.

If we, on the other hand, look at the rocks within the anorthosite area, the connection between them is also clear. There are all transitional links between anorthosite, anorthosite-gabbro, and basic segregations. The mineral assemblage is also the same. The large basic segregations contain the same mineral assemblage as the small spindle-shaped lenses marking the flow lines of the anorthosite.

As mentioned above, there is within the anorthosite area a certain distribution of the different rock types, the pure anorthosite lying in the interior while the dark minerals on an average increase towards the boundary.

The question is then whether there exists a genetic relation between the anorthosite and the mangeritic rocks around it. All the descriptions of anorthosites show that they occur in connection with granodioritic or syenitic rocks (J. Bugge 1943), but the genetic relation between them is not always clear.

In the area investigated here there are a lot of features that suggest the genetic relation between the rocks. Most of them have been mentioned above; here the most important will be more precisely stated.

There are never sharp borders, but transitional zones between the anorthosite and the surrounding rocks.

Within the anorthosite area zones of acid and intermediate rocks occur. The mineral assemblage of these is the same as that which is found in the surrounding rocks. The anorthosite massif Valafjell—Larsfonnfjell—Brekkenipo lies quite surrounded by mangeritic rocks, and must be regarded as an outlier in relation to the large anorthosite massif. It may therefore be said that anorthosite also occurs in the mangerite. It may be difficult to explain this fact without presuming an approximately contemporaneous formation. The differentiation may, however, have taken place over a long period of time, so that a considerable difference of age separates the first segregated and the last consolidated rocks.

A point also in favour of magmatic differentiation is the occurrence of gabbro "in statu nascendi" within the anorthosite. As mentioned this is a banded rock, and in the bands representatives of practically all rock types within the anorthosite kindred are found. The most reasonable explanation of these facts is that the magma has consolidated before the differentiation was completed.

The anorthosite rests in a bowl of mangeritic rocks. Both the anorthosite and the mangerite are foliated in the boundary zone. The foliation follows the boundary and is independent of the Caledonian direction. This is quite analogous to the conditions described by Balk (1931) from the Adirondacks for gabbro-anorthosite or gabbro-syenite. From the structural conditions he concludes that the gabbro must have sunk in relation to the anorthosite or to the syenite. Analogously we may here conclude that the anorthosite must have sunk in relation to the mangerites.

The area exhibits also in other ways several points of resemblance with the Adirondacks, e. g. the gradual zones of transition and the occurrence of shear zones in the anorthosite. This similarity is also a feature that favours the view that the rocks have been formed by magmatic differentiation. If we interpret the rocks as co-magmatic, the mangerite magma must be regarded as the residual or "mother-liquor" according to Balk, representing the composition of the magma after the segregation of the gabbros and the anorthosite. A consideration of the variations of the An % in the plagioclase in the various rocks supports this interpretation.

In the unaltered anorthosites the An % is c. 60 with values up to 63; towards the boundary it decreases to 40—50 %. However, since these rocks contain more epidote, the difference in An % is not necessarily due to the chemical composition, but may be caused by a difference in temperature during the consolidation. This applies to the inner part of the border zone, i. e. to the foliated anorthosite rocks. In one of these rocks it has been calculated how great the An % in the plagioclase would have been if all the lime had been absorbed by the plagioclase. The following equation has been employed (Vogt 1927):

3 plagioclase + 3 CaO (from the hornblende) → 3 clino-zoisite.

As an example is used an anorthosite-gabbro from Larsfonnfjell (vol.anal. No. 3, p. 165). The calculation shows that the An % would have increased from 50 to 65. This indicates that the difference in An % may be due to a difference in temperature.

In the mangerites the An % is 20—25; the jotunite has c. 27. The diorite transitional rocks have very variable An % in the plagioclase, forming a practically continuous series with values from below 30 to 50. But these rocks are quantitatively quite subordinate in relation to mangerites and anorthosite. Taking the distribution into account we get two pronounced maxima of the An % in the plagioclase, one at 20—25 corresponding to the mangerites, and one at 60—63 corresponding to the anorthosites.

Now we know that in a plagioclase melt plagioclase crystals of the composition An_{65} is in equilibrium with a melt of the composition An_{25}. The accordance between the theory and the conditions in the field is here very good. From what has been said it should follow that the rocks here must have been formed by magmatic differentiation. The problem then is how this differentiation has taken place.

This problem has been thoroughly discussed in geological literature, e. g. by Bowen 1917, Balk 1931, von Eckermann 1938, Buddington 1939. It will not be treated in all its aspects here. As mentioned above the rocks show several points of resemblance to those in the Adirondacks, and Balk's theories about the magmatic differentiation there may to a large extent be applied to the rocks of the anorthosite kindred in Vossestrand.

It is of little use to make assumptions as to the composition of the original magma as long as the entire area has not been investigated. Goldschmidt states his opinion in the following way: "Die Durchschnittszusammensetzung der sichtbaren Bergen—

Jotun-Gesteine dürfte ungefähr dem Jotun-norite entsprechen." Nothing more certain can be said as a result of the present investigations, nor about the first stages of the differentiation. With Balk we believe that the anorthosite and the anorthosite-gabbro have been segregated as crystal differentiates. This crystal paste has then moved along in the still liquid mother-liquor. As to the cause of the differentiation Balk says: "To judge from a large body of observations, the principal factor which has stimulated the differentiation of gabbros and anorthosites is the friction which the magma encounters as it moves along relatively stationary walls." And the mortar texture in the anorthosite should clearly prove that such movements really have taken place in the segregated crystals. It is, however, not always easy to decide what is due to protoclasis and what is due to cataclasis. In the chapter on the metamorphism, structures in the anorthosite will be mentioned which are certainly cataclastic. But otherwise the brecciation is so complete throughout the anorthosite massif, even in the most massive types, that it hardly can be interpreted as anything but protoclastic. It has been shown that the crushing in the anorthosite area must be older than the recrystallization in the boundary zones.

During the movement of the crystal paste a small portion of it — which now constitutes the anorthosite area between Brekkedalen and Uppheimdalen — must have separated from the bulk. Finally, then, the last remnants of the mother-liquor have been separated. The expression used by Bowen "squeezing out" is the most adequate term. The foliated mangeritic and acid zones that we find in the interior of the anothosite massif, must be regarded as remnants of the mother-liquor that has been squeezed out.

The structural conditions around the anorthosite massif show that the latter must have sunk in relation to the surrounding rocks. Above has been mentioned that the foliated boundary zone also constitutes a transitional zone to the surrounding mangeritic rocks.

This boundary zone is one of the most characteristic features of the anorthosite massif, and the question as to its formation one of the most important. Before attempting to give an answer to this question we will repeat the chief features of the profile section. Beginning at the bottom with the mangerites, there is from them a gradual transition to diorites. Above lies a zone of anorthosite and anorthosite-gabbro. This zone is distinguished by having a relatively low An % in the plagioclase and by containing much epidote. Besides the plagioclase is recrystallized to a large extent, and there is newly-formed hornblende. The recrystallization is most pronounced nearest to the border and decreases inwards (v. p. 142). Above this zone, again, lies more or less massive anorthosite, with higher An % and little or no epidote. The structure is here excessively crushed, protoclastic. The mangerites outside also show incipient mortar texture, but it is not so marked as in the anorthosite.

This marginal zone is here interpreted as a zone of reaction between the anorthosite and the surrounding mangerites, formed as the anorthosite was sinking relatively to the mother-liquor.

Another possibility is that it has been formed later; this will be treated in the chapter on the metamorphosis.

The fact that the mangeritic rocks around the massif exhibit incipient mortar texture indicates that they were not altogether liquid as the anorthosite was sinking. On the other hand, the structures around the massif indicate that they must have been plastic. It is most reasonable to presume that the mangerites began to consolidate during the sinking of the anorthosite. A regular plutone will have, during the consolidation, the highest temperatures in the interior and a zone of lower temperatures in the boundary zones. There is no reason to believe that the anorthosite massif should behave otherwise in this respect. The anorthosite massif with which we have to do here is distinguished by a relatively large content of dark minerals. It is not easy to give any figures, but I will estimate the content of dark minerals to c. 15 % on the average, or at any rate between 10 and 20 %. It may be mentioned here that Daly (1933) states 12 % dark minerals to be an average for the Voss—Sogn massif. My impression is that the anorthosite massif here described possesses a considerably higher content of dark minerals than the rest of the area.

Bowen (1917) discusses the relation between the content of dark minerals in the anorthosite and the corresponding content of melt in the "anorthosite-magma" or the crystal paste. He finds that an anorthosite with 2 % diopside may have contained 10 % melt, one with 5 % diopside 20 %, and one with 10 % diopside 35 % melt. The composition of the melt must lie on the boundary curve plagioclase-diopside in the triangular diagram albite-anorthite-diopside. Bowen further emphasizes that every igneous rock will pass, in the course of its consolidation, a stage where it is 80 % crystalline. Therefore it may, under otherwise similar conditions, assume the same structures as an anorthosite with 5 % diopside. The only difference generally is that the anorthosite is more coarse-grained and has protoclastic structure.

According to these considerations this "anorthosite-magma" should have contained in any case more than 15 % melt. And according to Bowen this quantity should be sufficient in order that the "anorthosite-magma" might be supposed to have been intruded. What is of greater interest in this connection, however, are the conditions of the temperature. With such a great content of melt, it is not unreasonable in any way to presume that the conditions of the temperature would have been similar to those of a normal plutone, i. e. with the highest temperatures in the interior and the lowest towards the boundary.

It has been mentioned that an extensive recrystallization has taken place in the boundary regions, most pronounced at the extremities. Farther towards the interior partial recrystallization is common. Here there are two generations of plagioclase grains, some large brecciated ones in a finely recrystallized ground mass. As mentioned above the An % is the same in the two generations. This shows that the whole process must have covered a long time.

The dioritic rocks, situated between the anorthosite-gabbro and the mangerites, are typical transitional rocks. The dark mineral is a more or less transformed pyroxene, often with a clear reaction zone. It may also be hornblende only. The plagioclase is most frequently recrystallized, An % 20—25 is common. Potash feldspar is found in the rocks situated nearest to the mangerites, the content decreasing towards the anorthosite-gabbros.

Viewed in connection with their mode of occurrence these rocks are most naturally interpreted as zones of reaction between the anorthosite rocks and the mangerites. The question then is at what stage this reaction has taken place. In the mangeritic rocks, as mentioned above, the consolidation has started under T- and P-conditions corresponding to the granulite-facies. During the consolidation T and P have decreased so that the final conditions were most like the hornblende-gabbro-facies. The uralitization of the pyroxene,

e. g., showed this transition. Uralites of the same type are also found in the dioritic links of transition, which indicates that they must have been formed at an earlier stage. The steady increase of the An content of the plagioclase shows that the reaction must have extended over a long period. The plagioclase traverses here a continuous series of reaction, from labradorite to oligoclase; in the dark minerals we get a corresponding series, from pyroxene over uralite to hornblende. Biotite is also very common in the dioritic transitional links and in the mangerites. This is in accordance with Bowen's principle of reaction.

The reaction zone extends some distance into the anorthosite, generally c. 100—200 m. Within this zone traces of potassium metasomatism have been found. Anorthosite and anorthosite-gabbro may contain patches filled with sericite. This may be so common that it can be explained in no other way than by a supply of potassium. It must be supposed that this potassium originates in the mother-liquor.

Finally some remarks should be added about the age relations between the various links. The anorthosite rocks are the oldest. It is difficult to state anything about the age relation between the links here, but their mode of occurrence hardly indicates any great difference of age between them. The mangeritic rocks are younger than these, but here, too, it is impossible to say anything about the age relation between the various links. Pegmatite dikes and silexite traverse all the preseeding links, and must be younger than the latter. As mentioned above, the pegmatite dikes cross the flow lines of the anorthosite. Silexite and pegmatite often occurs side by side, but as it has never been observed that they cross each other, their relative ages cannot be determined. Finally diabase dikes are younger than the silexite.

It must here be presumed that the anorthosite rocks have been segregated as crystal differentiates, the intermediate and acid links have consolidated as normal magmatic rocks, the pegmatites have been formed in the pegmatite phase, and the silexites are aqueous igneous or hydrothermal formations. This is, it must be said, a very incomplete scheme of differentiation, but it must be borne in mind that the area of unaltered igneous rocks is very limited, the metamorphism having obliterated the majority of the original features over the greater part of the area.

THE METAMORPHISM. GENERAL SURVEY

As mentioned above it is biotite-gneisses of parallel structure that constitute the majority of the rocks of the quadrangle. The connection between these gneisses and the described fresh mangerites, jotunites, etc. is clear. The metamorphism has not been so complete everywhere, and the various stages may be followed from the fresh igneous rocks in the hornblende-gabbro-facies to the schistose-biotite-gneisses in the epidote-amphibolite-facies. Kolderup (1940) uses the following terminology about the metamorphic rocks of the anorthosite kindred: "If the rocks have parallel texture, and the microperthite is replaced by microcline, the biotite rocks may be termed mangerite-gneisses, the hornblende rocks amphibolites."

According to this statement practically all the gneisses here are mangerite-gneisses. Since it is possible, however, in some cases to decide whether the rock has been a jotunite, a mangerite, a birkremite, the terminology jotunite-gneiss, birkremite-gneiss, etc. is used. Macroscopically these various types cannot be distinguished from each other. On the map, therefore, mangerite-gneisses have been marked over the entire area.

The metamorphism must be regarded as a result of a thrusting. A number of measurements of gliding striæ show that the rock has been thrust from NW to SE. Other proofs of the thrusting are an extensive heaping of the rocks along the thrust front, and a crushing, partly a mylonitization, on both sides of the thrust plane. Also otherwise in the field a great many zones of brecciated rocks are found having a clear relation to the thrust direction. This will be further mentioned in the chapter on tectonics.

Broadly speaking we may distinguish between a mechanical and a chemical effect of the metamorphism. The rocks have been exposed to an extensive crushing, and in single cases pure mylonites and breccias have been formed. But most common are cataclastic 'eyed' gneisses with 'eyes' of potash feldspar or plagioclase.

There are definite zones where the crushing has been particularly strong, but it can be traced over the entire area. Even the freshest types show incipient brecciation, but as mentioned above, it is here difficult to decide what is due to proto- and what to cataclasis.

Recrystallization is very common in the regular gneiss types. It is first of all quartz that has been new-formed. The recrystallization must have taken place contemporaneously with or after the crushing, the new-formed minerals being little or not influenced by the mechanical metamorphism. The new-formed quartz forms small veins and bands parallel to the schistosity. Next to quartz biotite is the most common of the new-formed minerals. It marks the plane of schistosity, and often it may be seen bending round the feldspar-porphyroclasts. Other new-formed minerals are chlorite, epidote, and in single cases, calcite. Feldspar is seldom or never new-formed, but a transformation from microperthite over microcline-microperthite to microcline is usual, as is the albitization of the microcline.

A certain migration of material has also taken place. The plagioclase in the gneisses is never fresh, but more or less filled with sericite and epidote. Often the quantity of sericite may be so great that it can only be explained by a supply of potash. The formation of potash feldspar, and both these phenomena are exceedingly the hornblende, a process which also requires a supply of potash. On the other hand potash is liberated by the albitization and seritization of potash feldspar, and both these phenomena are exceedingly common. Quantitative investigations of this have not been undertaken, but there is at any rate a possibility that the two processes may equal each other, and that it is not necessary to presume a supply of potash from without.

THE METAMORPHISM IN THE ANORTHOSITE AREA

As mentioned above, the rocks of the anorthosite area are little influenced by the metamorphism, but nevertheless its effect is spotted. There is everywhere an extensive crushing. The difficulty is to decide what is proto- and what is cataclasis. That movements have occurred after the consolidation is certain; that is *i. a.* shown by a number of small faults in the dark bands. Faults have also been observed in the shear zones.

The foldings have been mentioned above; they are large, wide folds that have been formed after the consolidation, but appear to have a connection with it, the folding-axes having a certain relation to the borders. They have, at any rate, no connection with the later Caledonian direction.

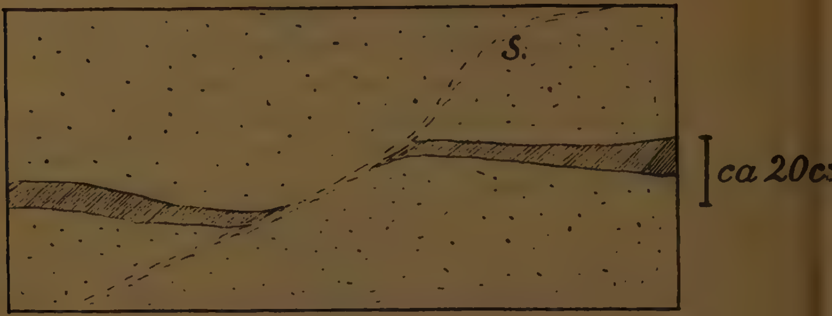


Fig. 18. Fault in dark band, Brekkenipo. S=shear zone.

Undoubted relation to the Caledonian thrust direction have, on the other hand, a number of crushing zones. Such zones occur also in the gneiss area, but they are most common in the anorthosite area.

On the structure map the chief crushing zones are marked. They can also be seen on a topographical map as they are most frequently indicated by the shape of the contours. This appears particularly clearly on Brekkenipo.

The rocks of these zones may be breccias or mylonites, but not always the crushing has gone so far. It is seen on the map that these crushing zones are particularly common in the boundary areas. The direction of these zones is NE—SW. On the NW and SE sides of the massif they follow the boundary roughly, but locally they cross it. The crushed zones on Brekkenipo have the same direction. One of these runs in the border area between the anorthosite and the gabbro rock in Hodnardalen.

It is evident that these rocks have been formed after the *mise en place* of the anorthosite, which appears *i. a.* from the fact that they may cross the boundaries. On the south side of Valafjell this appears clearly. If the crushing had taken place contemporaneously with the sinking of the anorthosite, it might have been expected that it had been equally strongly developed all around the massif. On the east side of the massif, on the slope down to Brekkedalen no continuous crushing zone has been found, the crushing having only been observed in a few small localities. Crushing zones also occur in the gneiss area running NE—SW. All this indicates that the crushing zones are Caledonian and have connection with the thrusting.

Similar phenomena connected with thrusting are well known. When a massif is being thrust forwards a crushing zone usually will arise in front of it. Such zones are also found here on the border between the gneisses and the mica schist. If the resistance to the thrusting becomes too strong, it may be imagined that the forces will get released by the formation of a crushing zone behind. This may be repeated later on, and in this manner we may get a succession of parallel crushing zones. These zones will form a definite angle with the thrust direction; here they are approximately vertical to it.

The fact that these crushing zones are more widely distributed within the anorthosite area than outside, shows that the rocks have reacted differently to the metamorphism. Outside the anorthosite area the metamorphism has been more complete, the crushing is as it were more evenly distributed in the rocks, the recrystallization is at the same time more common, and the alteration is so extensive that is is often difficult to recognize the original rock. The anorthosite rocks have been more capable of resistance, the crushing being limited to definite zones while the rest of the area is rather uninfluenced. As a boundary always represents a weak zone, it is reasonable that these zones have been formed first of all in the boundary areas. In this connection it may also be borne in mind that the unaltered intermediate and acid rocks that are found in the area, are situated in the vicinity of the anorthosite. This has itself resisted the influences of the metamorphism, and also protected the rocks near by. One may be tempted to say that the latter have lain under the lee of the anorthosite massif.

The recrystallization is completely lacking in the interior of the anorthosite massif, but it is common in the boundary zones. The recrystallization here has been mentioned above; it was interpreted as having taken place before the metamorphism, during the *mise en place* of the anorthosite. Another possibility which cannot be disregarded is that the recrystallization may have taken place during the metamorphism. The entire marginal zone around the anorthosite massif must then be interpreted as a metamorphic formation.

As mentioned above, the majority of the mangeritic rocks of the field have been transformed to acid biotite-gneisses during the metamorphism. This transformation must have been caused by solutions circulating in the rocks. We may then imagine the formation of the boundary zone in the following way: These solutions have been capable of penetrating some way into the border areas of the anorthosite massif. The metamorphism has not been so complete that the anorthosite has assumed the Caledonian schistosity. But it may be imagined that the solutions have penetrated along the previous "magmatic" structures, and have effected a recrystallization of the rocks here. The foliation along the boundary must at any rate have been formed during the *mise en place* of the anorthosite.

There are several objections to this theory. In the first place: All the metamorphic rocks elsewhere in the field, both the amphibolites and mangerite-gneisses are in the epidote-amphibolite-facies, while the rocks in the boundary areas of the anorthosite belong, as mentioned above, to the hornblende-gabbro-facies, or if they are interpreted as metamorphic to the amphibolite-facies. This difference of facies is difficult to explain by this theory.

Upon the whole it seems to me that the majority of the facts are in favour of interpreting the boundary zone as pre-Caledonian or early-Caledonian formed during the *mise en place* of the anorthosite.

It is brought out by the map that the mylonite zones in the vicinity of the anorthosite massif separate the Caledonian structures in the gneisses from the older structures in the anorthosite massif. These areas have been investigated in details in order to find, if possible, traversing structures, but no such have been found.

THE METAMORPHISM OUTSIDE THE ANORTHOSITE AREA

A Survey.

In these rocks the metamorphism has been far more complete. Over the greater part of the area we find amphibolites and mangerite-gneisses in the epidote-amphibolite-facies. The rocks have not been equally strongly transformed everywhere in the field, and a certain regional distribution of the metamorphism can be pointed out.

The least altered rocks are found in the vicinity of the anorthosite massif, especially on the south-east side of it. The only exception to this rule is a small area of mangerite and mangerite-syenite at Ulvund near Myrkdalsvatn.

As mentioned in the general survey a distinction may be made between mechanical or chemical metamorphism or between crushing and recrystallization. Such a distinction is of course purely formal, and it serves only to facilitate the exposition. In reality the two processes will always overlap.

A purely mechanical deformation would hardly have occurred in such a region of mountain ranges as this.

In the breccias the mechanical deformations are thoroughly predominant, and here the recrystallizations are limited to the joints. In the dusty ground mass of the mylonites there will always be recrystallization, even though it usually is difficult to demonstrate.

The rocks of predominantly mechanical deformation are found first of all in the boundary zones of the mica schist. The crushing is most pronounced along the thrust fronts where breccias and mylonites have been formed. Strongly mechanical deformed rocks are also found in local crushing zones in the interior of the gneiss area. These zones are, as a rule, parallel to the thrust front.

In a metamorphism that is predominantly mechanical, the size of the grains will decrease with increasing metamorphism. Roughly speaking the size of the grains decreases here towards the west. Especially in the region of Skreia and Mykjedalsvatn the rocks are perfectly fine-grained, and at the same time they are markedly schistose. Without having seen the more coarse-grained transitional types it would not have been possible to decide whether these rocks belong to the mangerite-gneisses. Under the microscope porphyroclasts of feldspar in a fine-grained ground mass are seen. These porphyroclasts in some cases consist of the characteristic microperthite of the anorthosite kindred, but often this has passed into microcline, and then it is very difficult to decide what sort of rock it has been originally. West of Skreia and Mykjedalsvatn the mangerite-gneisses border on schists rich in quartz that have the same strike and dip. These being also very fine-grained it is difficult to distinguish them from the mangerite-gneisses even under the microscope. So far I have not been able to fix any divisions between these rocks.

Nor is it a rule without exceptions, that the size of the grains decreases towards the boundary. Small areas of fine-grained rocks are found otherwise in the field, too, but roughly speaking the coarse-grained 'eyed' gneisses prevail over the greater part of the area. These are porphyroclastic 'eyed' gneisses where there has been an extensive recrystallization along the planes of schistosity.

As mentioned above the terms mangerite-gneisses and amphibolites are used about these rocks, but there also is a need for a classification according to the degree of metamorphism. It is then natural to classify them according to the degree of crushing and recrystallization. In the course of time many suggestions of such a classification have been made, but the terminology has been very changing and not at all unambiguous. In 1935 Waters and Campbell published a paper on mylonites where they endeavoured to introduce order into the terminology. They gathered the most important of the rocks in a clear and simple scheme, which is reproduced here:

Crushing		Recrystallization	Fusion
Near surface (Material incoherent or cemented by later secondary processes. Pressure and temperature low).	*At depth* (Material remains coherent throughout ruptural deformation. Pressure moderate or strong, temperature low).	(Ruptural deformation occurs under sufficient pressure and at a high enough temperature to bring about extensive recrystallization).	(Temperature during deformation high enough to result in partial or complete fusion).
Gouge and finely pulverized fault breccia.	Ultramylonite. Some Flinty crush Rock.	Hartschiefer.	Pseudotachylyte. Some Flinty crush Rock.
	Mylonite.	Blastomylonite.	
		Mylonitegneiss.	
Fault breccia.	Protomylonite.		

Increase in grain size ↓

Increase in pressure and temperature →

As to the definitions of the various types reference is made to the mentioned paper by Waters and Campbell.

The majority of the metamorphic rocks belong to the group mylonite-gneisses. This is just what might have been expected in advance. As the area belongs to the Caledonian Mountain Range, it is but reasonable that the recrystallization makes itself strongly felt. Mylonites, breccias, and protomylonites also occur, which are, however, not so common.

Description of the Gneiss Rocks.

The microscopical examination reveals an obvious connection between the igneous rocks and the gneisses. The metamorphism has not been equally strong everywhere, and by finding representatives of the various stages of the metamorphism the gradual development may be followed from fresh igneous rocks over 'eyed' gneisses to fine-grained gneisses, or in the case of a prevailing mechanical deformation, to breccias and mylonites. In order to illustrate this transition a number of thin sections have been photographed (Figs. 19—23). The samples are taken from the different parts of the field and are only meant to show the different stages of the metamorphism, irrespectively of the regional distribution.

Fig. 19 shows a mangerite-syenite from Ulvund. The rock is comparatively fresh, and its dark mineral is hornblende. On the picture a large microperthite grain is seen. It exhibits incipient mortar texture, is traversed by a lot of joints where there is crushed, partly also recrystallized, material.

By continuous crushing the joints will grow at the expense of the original mineral grain, and the fine-grained material will increase. Whether both recrystallization and crushing occur side by side will depend upon the pressure and the temperature conditions during the metamorphism. Besides it is also dependent upon the composition of the rocks. It is a general experience that dark minerals will recrystallize more easily, while e. g. feldspar under equal conditions will exhibit cataclastic texture. Likewise quartz is less capable of resistance to crushing than feldspar, and it is also one of the first minerals to become new-formed. These facts are excellently illustrated in the gneiss rocks of this area. We most usually find porphyroclast of feldspar while quartz and the dark minerals have been recrystallized. Harker (1932) characterizes this as follows: "In crystalline rocks other than calcareous the most freely soluble mineral is usually quartz. When such a rock as a granite or gneiss is subjected to great shearing stress, the quartz elements, by solution at some point

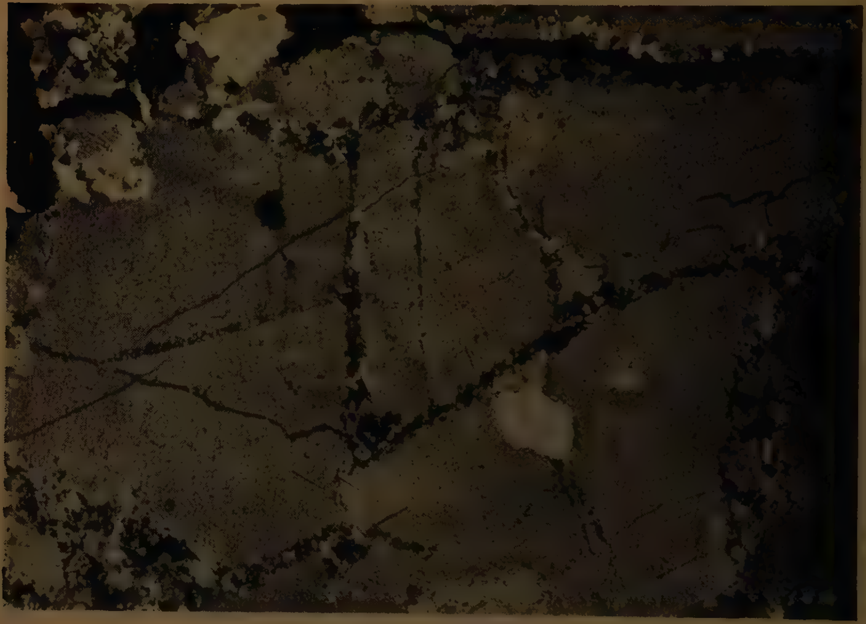


Fig. 19. × 10.

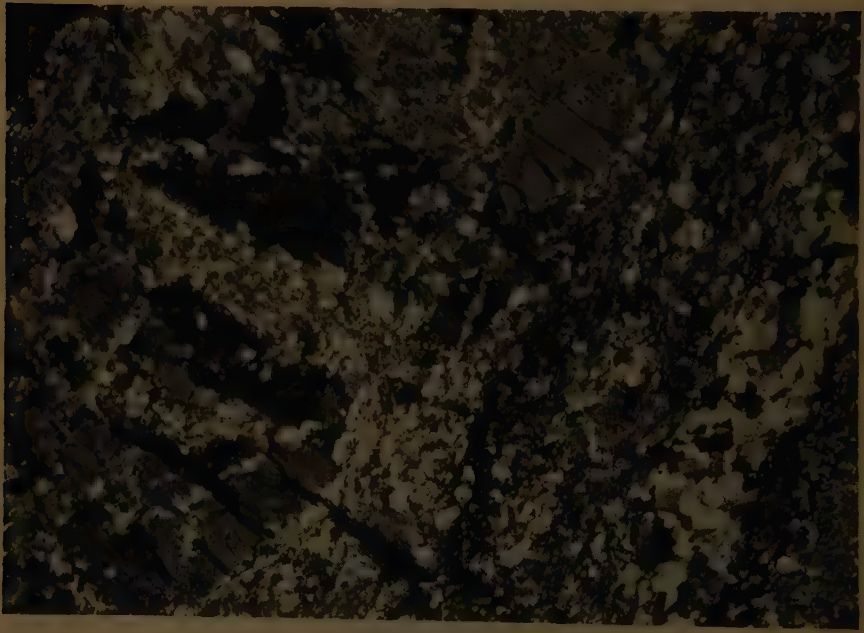


Fig. 20. × 10.

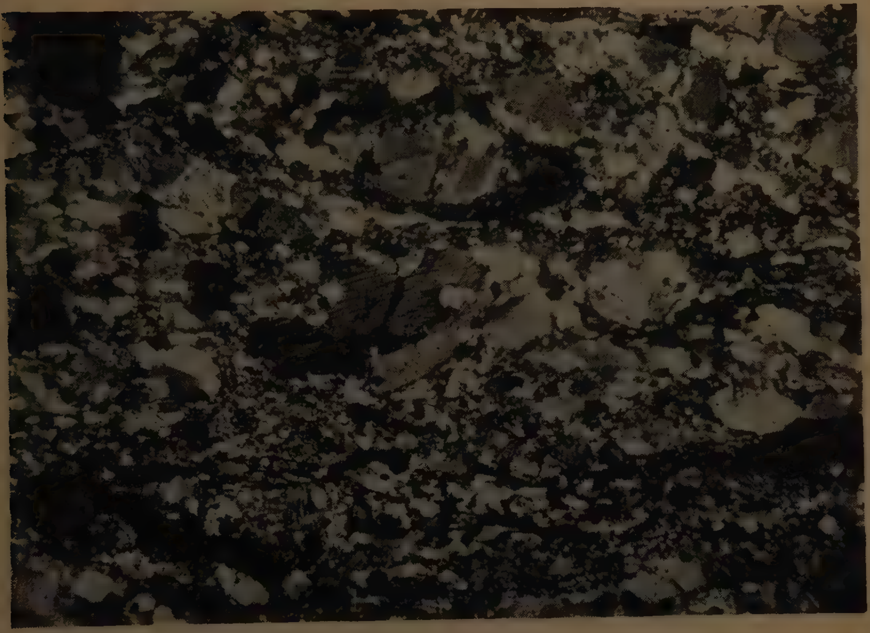


Fig. 21. × 10.

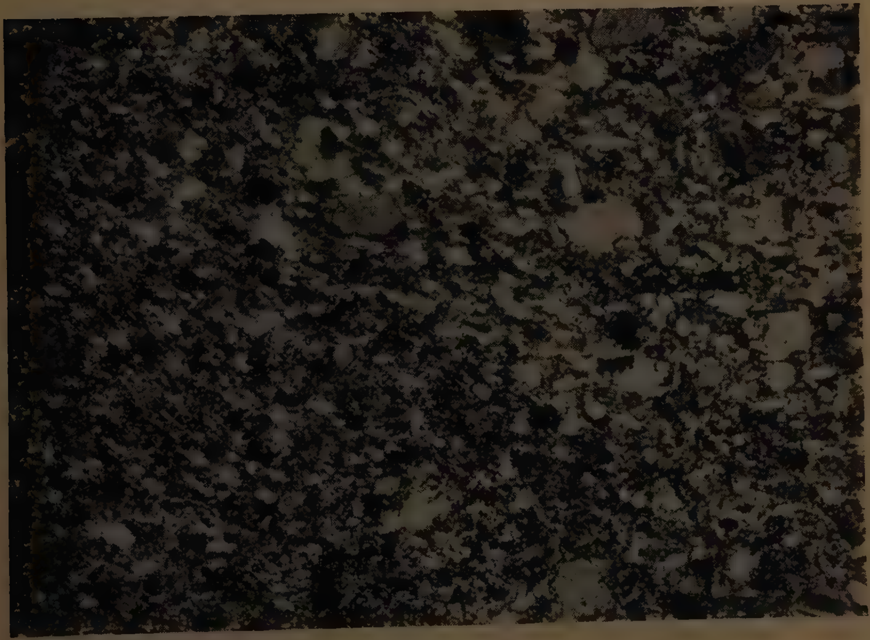


Fig. 22. × 10.

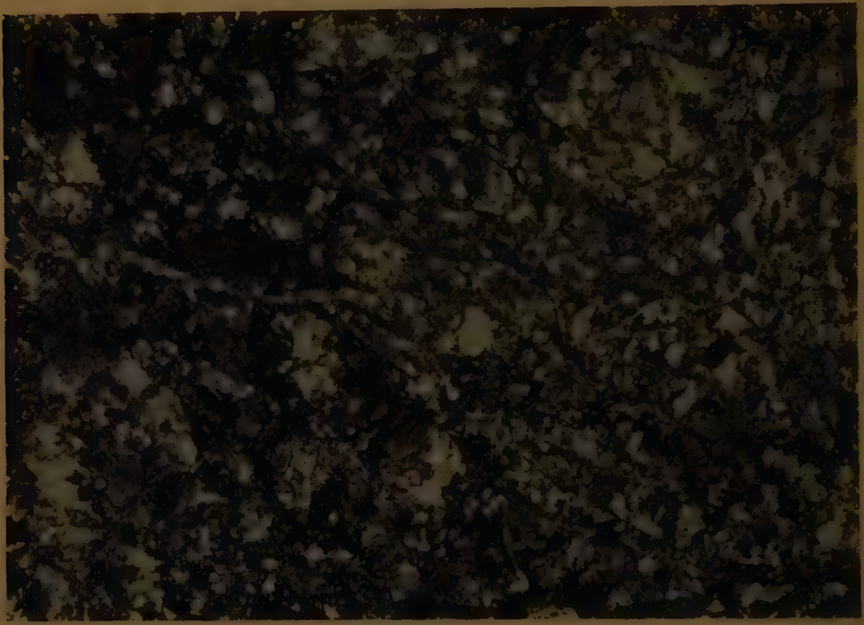


Fig. 23. × 10.

and concurrent addition at others, are capable of gradually changing shape, always in the sense of yielding to and relieving the internal stress. This readjustment, co-operating with cataclastic effects in the other minerals, is a factor of importance in the quasi-plastic deformation of the rocks.«

That also the dark minerals are more susceptible to recrystallization than e. g. feldspar, is a general experience. About this Waters and Campbell (1935) say: "Evidently under the pressure-temperature conditions that result in the formation of a mylonite from quartzite or graywacke, basalts and diabases are partially recrystallized, and serpentine is entirely recrystallized, and converted into schist.«

Fig. 20 shows a regular 'eyed' gneiss of the most common type in the gneiss area. Here the crushing has gone considerably further. The joints have widened so that the originally large microperthite grain has been almost completely devoured. It is worth noting the characteristic lenticular form of the feldspar-porphyroclasts with the longest axis parallel to the schistosity. The rock is a typical 'eyed' gneiss: macroscopically the whole crushed microperthite grain, i. e. the entire left triangle of the picture, forms an 'eye'. To the extreme right of the picture new-formed quartz and biotite are seen marking the schistosity. Further a number of cross joints filled with chlorite are seen. This is, according to the degree of metamorphism, a typical mylonite-gneiss.

Fig. 21 shows a rock where the metamorphism has been more thorough. Here it is no longer possible to perceive the outlines of the originally large grains. The feldspar-porphyroclasts are smaller, the microperthite predominates, but a few grains are microclinized. The 'eyed' structure is typical still. The ground mass has been exposed to an extensive recrystallization.

Fig. 22 shows a sample taken just beyond the border of the mica schist. It belongs to the most fine-grained gneiss types. The porphyroclasts are here small and few and the microperthite spindles seldom preserved. When the ground mass is so fine-grained it is very difficult to decide whether a recrystallization has taken place. In this case it may be seen that some of the grains in the ground mass have followed the Riecke principle, lying with their longest axis parallel to the plane of schistosity. Consequently these grains must have been new-formed.

The last picture of this series (Fig. 23) is taken from the vicinity of the thrust front at Ulvund. Macroscopically the rock is quite compact, and has a bluish tinge suggestive of blue quartz. Under the microscope some few porphyroclasts are seen in a dusty ground mass, but there are no signs of recrystallization. This rock must be designated as a typical mylonite. On the photo the fluxion texture is clearly seen.

The most important types of the gneiss rocks occurring in the field are represented in this series. Macroscopically the gneisses exhibit great variations, and only a microscopical examination of a lot of samples made it possible to ascertain that we have to do with rocks of the anorthosite kindred.

We will then pass on to a further description of the various gneiss rocks.

Amphibolite. Amphibolites are rather common, but they never form any large, independent areas. They are found in small occurrences among the intermediate and acid gneisses, and are connected with them by gradual transitions. They have also the same strike and dip as the gneisses. Samples from a total of 10 localities have been examined. They exhibit, macroscopically, small variations. They are as a rule medium- or fine-grained rocks with a typical greenish tinge. They show no visible signs of crushing. A microscopical examination shows that the mineral assemblage is roughly the same, only rather small variations occurring. Nor under the microscope do these rocks exhibit any signs of crushing. But there is on the other hand an extensive new-formation of minerals.

The chief minerals are hornblende and plagioclase. The hornblende is the common bluish green one often coloured dark by a pigment. The pigmentation is stronger in the middle of the grain, which often may be completely dark while the marginal zone is as a rule not coloured. The pleochroism is the usual one, $\gamma =$ bluish green, $\beta =$ green, $\alpha =$ yellowish green. Indices of refraction of the hornblende has been determined in three of the samples.

1) By the river to Svarttjern: $\gamma_{Na} \sim 1.678$, $\alpha_{Na} \sim 1.658$ c/γ 15°
2 V ∼75°
2) North-west of Geitafjell: $\alpha_{Na} \sim 1.664$ c/γ 19°
3) At Giljarhus: $\alpha_{Na} \sim 1.664$ c/γ 18°

According to Kunitz' tables this corresponds to syntagmatite-hornblendes with 47, 55, and 55 % Fe-compound respectively. In the other samples the angle of extinction only has been measured, varying from 15° to 21°.

The plagioclase is never fresh. Most frequently it is quite filled with a pasty mass of clino-zoisite, muscovite and epidote, to such an extent that it is impossible to recognize the original plagioclase grains. Measurements of the index of refraction of a sample gave the result: $\alpha_{Na} \sim 1.530$. ɔ: 7—8 % An, but generally the plagioclase is so saussuritized that it is impossible to measure the index of refraction.

Quartz, muscovite — usually in the shape of sericite — and a mineral of the clino-zoisite—epidote series have been new-formed at the expense of the plagioclase. The quartz occurs either as single grains equally scattered throughout the saussuritized ground mass, or it may form bands and veins, these then marking the schistosity. The quartz grains are, as a rule, only faintly undulating.

The epidote-mineral is found most frequently in small grains together with sericite in the plagioclase grains, but it also occurs as rather large grains. The content of iron silicate varies, sometimes amounting to 18—20 %, determined according to the interference colours in a normal section.

Biotite is found together with the hornblende and is, as a rule, an alteration product of it. It is brown. In a few samples it possesses inclusions of fine rutile needles, sagenite. The transformation may have gone so far that it prevails over the hornblende.

Ilmenite is always present. It is surrounded by a coating of titanite or leucoxene.

Potash feldspar occurs in one of the samples, and is here a microperthite. As a rule the grains are albitized at the edges.

Apatite is common. A little garnet may occur. One sample contains a little calcite as a transformation product of the plagioclase.

These rocks are regarded as transformed gabbros, or jotunites if they contain more than 12 % potash feldspar. A feature common to them all is that they show no signs of crushing, but, on the contrary, extensive recrystallization of the minerals, hereby distinguishing themselves from mangerite-gneisses. This difference must be due to the rocks having reacted differently to the metamorphism. As they contain a lot of dark minerals it is reasonable that recrystallization prevails.

Mangerite-gneisses. The acid and intermediate gneisses comprise the greater part of the investigated area. A total of 60 thin sections of these rocks have been examined, so great a number having been necessary because the rocks, macroscopically, may look so quite different. Most common is an 'eyed' gneiss with 'eyes' of reddish feldspar. Regular banded gneisses are also found. The feldspar may be both red and white here. Very common are also a lot of fine-grained types varying in colour from white and greyish to quite dark. Besides, breccias, mylonites, and other crushed rocks are common, these having often a reddish tinge. Finally, also banded gneisses with fluxion texture are found. By a macroscopical examination only, it is difficult to recognize the relation between the various types, but under the microscope they appear to have, roughly speaking, the same mineral assemblage. Structurally they exhibit also greater similarity under the microscope than in the field. Crushing is typical, and the difference of grain size is conditioned by a different degree of crushing, the size decreasing with increasing crushing. Nearly always minerals have been new-formed, but the proportion between recrystallization and crushing varies greatly. Generally, crushing prevails, but in the banded gneisses with fluxion structure recrystallization is practically sovereign.

The chief minerals are: Acid plagioclase, potash feldspar, quartz, and biotite. The quantitative relation between the minerals varies in the same manner as in the igneous rocks. On the map all the intermediate and acid gneisses are indicated by one symbol, viz. as mangerite-gneisses, but under the microscope it has been possible to separate a few types. Theoretically a distinction should have been possible according to the quantitative relations of the minerals, in jotunite-, mangerite-, birkremite-gneiss, etc. Because of the rather strong metamorphism, however, it is impossible to decide to which family the rock originally has belonged. A certain migration of materials has also taken place during the metamorphism. All the gneisses abound in new-formed quartz which may originate by the saussuritization of the plagioclase of the rock, — it is always more or less altered, — but it may also have been supplied.

As to the mineral composition the various gneiss types exhibit no regional distribution, but are intimately intermixed over the entire area. Among the fresh igneous rocks the mangerites are without comparison the most dominating rock. Among the gneisses, rocks with more than 12 % quartz prevail. In the majority of the rocks potash feldspar prevails over plagioclase. Of course it may be imagined that the rocks here originally have contained more quartz than is the case in the fresh rocks now preserved, but it is equally reasonable that these rocks have been predominantly mangeritic, and that the quartz has been supplied later. It is then most natural to think that it has been formed by an incipient granitization during the metamorphism. This quartz occurs in a typical way in bands parallel to the schistosity, and has, as a rule, only faintly undulating extinction. With the microscopy a distinction can be made between rocks with prevailing potash feldspar and rocks with prevailing plagioclase; mangerite- and jotunite-gneisses respectively. The mineral assemblage is otherwise the same, and macroscopically it is impossible to distinguish one from the other.

The chief minerals are: Plagioclase, albite or acid oligoclase, potash feldspar, microperthite and microcline, quartz and biotite. In subordinate amounts: Hornblende, an epidote-mineral, chlorite, apatite, muscovite, garnet, and calcite. In accessory amounts: Titanite and zircon.

The plagioclase is seldom fresh. It is more or less saussuritized. As a rule the transformation is most pronounced in the interior of the grain, while the marginal zone is exempt from inclusions. The microlites have most frequently crystal outlines. Sericite and a mineral of the clino-zoisite—epidote series have been determined.

A great many determinations of the indices of refraction in the plagioclases have been made, totally 25. The An % varies from 6 to 22, but there is a clear maximum at 15—16 %, about one half of the examined samples falling within this interval. On an average the An % decreases with increasing quartz, so that the granitic gneisses, as a rule, contain albite. There are, however, exceptions to this rule. As an example a microcline-gneiss may be mentioned here from the region of Skreia. The plagioclase has here an An % of 22, $\alpha_{Na} \sim 1.540$. It contains decidedly more than 12 % quartz. Here the quartz does not look like having been new-formed, and it is reasonable to regard this rock as an altered charnockite.

Albitization of potash feldspar is common, and in some of the samples the bulk of the plagioclase has been formed in this way.

Twin lamellae are common in the plagioclase, usually according to the albite law, but percline twins also occur. The grains often are twisted or crushed, and undulating extinction is common. The grains never exhibit crystal outlines, but have the characteristic lenticular form that is common in cataclastic rocks.

Potash-feldspar. Both microperthite, microcline-microperthite and microcline occur. More seldom grains occur that have neither twin lamellae nor perthite spindles. The microperthite does not contain the large regular spindles common in the unmetamorphic types. During the metamorphism the microperthite has undergone a gradual alteration. As a rule the size of the spindles decreases, being pressed together, thus often becoming so slender that they are difficult to distinguish. It happens, however, that the spindles increase during the metamorphism, at the same time as new spindles most frequently are formed, too. These are large and irregular, and clearly formed by replacement. Often two generations of spindles may be seen in one and the same grain. With increasing metamorphism the grains generally become microclinized. It is common to see microperthite grains with a marginal zone that is transformed to microcline. Microcline-microperthite generally occurs as a transitional link between the two types. As the last stage of the process we then get an albitization of the microcline. This process also begins at the edge of the grain; large, irregular veins and tongues of albite penetrate from the edge, and finally transform the entire grain.

The potash feldspar is not so inclined to sericitization as the plagioclase. A characteristic feature of the microperthite is that the spindles are always altered earlier than the ground mass. Upon the whole the potash feldspar only seldom shows signs of alteration to sericite. Also the potash feldspar grains show the same lenticular shape as described for the plagioclase.

Quartz is found in all the intermediate and acid gneisses. As mentioned above it is difficult to decide how much is new-formed, particularly when the quartz grains are evenly distributed in the rock. It very often forms veins or bands parallel to the schistosity. These are clearly new-formed. During the crushing quartz is formed on joints in the other mineral grains. This is particularly typical in the feldspar grains (Fig. 19). Even the new-formed quartz exhibits, as a rule, undulating extinction.

Among the dark minerals the biotite is entirely predominating. Pyroxene is not found, and a little hornblende occurs only seldom in connection with the biotite.

The biotite is green with strong pleochroism: $\gamma = \beta =$ green, $\alpha =$ light green to yellowish green. It may also be brownish green, but purely brown biotite does not occur in the gneiss area. The biotite is clearly recrystallized, and it is gathered in definite planes which are the planes of schistosity in the rock. It often may be seen how it bends around the feldspar porphyroclasts, and how it clearly has been formed later than these. It often has been partly altered to chlorite.

Chlorite is otherwise found preferably on joints. Most frequently it is a penninite with characteristic lavender-blue interference colours. Segregations of small ore grains are common in the chlorite, especially where the mechanical deformation has been strong, plenty of chlorite has been formed.

An epidote-mineral is nearly always present both as microlites in the plagioclase and as rather large grains otherwise in the rock. Most frequently it is clearly secondary, but in some cases it looks like having been formed at an early stage. As an example an orthite-epidote from a granitic gneiss at Skjervheimstølen will be mentioned. Goldschmidt (1916) says about the biotite-granites of the Bergen—Jotunheimen tribe that they distinguish themselves by containing orthite-epidote.

Orthite-epidote is common here, especially in the acid gneisses. They often may have a considerable size, and in the sample mentioned 0.84 × 0.24 mm has been measured. The grain has a lengthened prismatic shape. The regular form and the size give reason to suppose that it may be primary. Besides orthite-epidote there is plenty of common epidote. The content of Fe-epidote is usually about 20 %. The measurements have been performed by means of interference coulours in a normal section. Winchell's tables have been used, but they give no accurate results. A total of 12 measurements have been performed that all give values from 15—23 % Fe-compound. Pure clino-zoisite is rare. Some of the microlites in the plagioclase show low interference colours, but since the microlites usually are very small it is not possible to state anything about the orientation.

Ore-minerals are always present. Most common is ilmenite surrounded by a coating of titanite or leucoxene. The width of this coating changes. If it is small it consists most frequently of titanite. The leucoxene coatings are usually broader. The coating often grows at the expense of the ore core, to such an extent that the latter disappears entirely. The formation of leucoxene increases with increasing metamorphism. The freshest types contain ilmenite and titanite.

Titanite is not only found as a marginal growth around ilmenite, but occurs in independent grains as well. Most usually the grains are fairly small, and it may then be difficult to distinguish them from zircon which occurs in the same manner.

Muscovite occurs together with biotite, most frequently quite subordinate in relation to the latter, but some rocks are found containing much muscovite. As an example a dark medium-grained rock at Flisramtjern may be mentioned, the chief constituents of which are microcline and muscovite. Under the microscope large grains of plagioclase and microcline are seen. They are crushed and traversed by joints where quartz and muscovite have been formed. The ground mass between the feldspar grains consists essentially of muscovite. The plagioclase is relatively fresh: $\alpha_{Na} \leqq 1.536$, ɔ: c. 16 % An. When so much muscovite is found, a supply of potash must be supposed to have taken place.

As mentioned above sericite is found as microlites in plagioclase.

Garnet does not belong to the usual mineral assemblage of the gneisses. It may be found in accessory amounts in small rounded grains. In rare cases the rock may contain somewhat more garnet, occurring then in round grains surrounded by green biotite. It has not been possible to decide whether these garnets are relics or newformed during the metamorphism.

Apatite is common. In the jotunite-gneisses it is always present. Most frequently the grains are small and rounded and rarely exhibit crystal outline.

Calcite is rare. Where it is found it is secondary. It occurs most frequently near by the plagioclase grains, and has been formed at their expense.

This description comprises the normal gneiss types. Some rocks have a mineral assemblage deviating from the normal one; on account of this they will be given a separate description.

Gneisses rich in epidote. In the western slope of Geitafjellet close by the silexite a greenish gneiss rock occurs. It may be quite fine-grained, but more coarse-grained types are also found. In these quartz, epidote, and a reddish feldspar may be distinguished macroscopically.

A volumetrical analysis of this rock has been performed. Volume analyses have otherwise not been made for the gneisses. Generally the ground mass is so crushed and the minerals so altered that this cannot possibly be done. Here also there is a fine-grained ground mass consisting of albite and potash feldspar. In the integration all the feldspar is taken as a whole. We then get the following mineral composition: 56.80 % quartz, 24.80 % feldspar, 10.80 % epidote, 5.01 % chlorite, 1.71 % ore, and 0.88 % titanite.

The quartz forms broad bands parallel to the schistosity. It is not impossible that quartz has been supplied during the advance of the silexite, and the abnormally high percentage of quartz may be due to this fact.

The epidote is poikilitic and is found in large accumulations. It constitutes the majority of the dark minerals, and is clearly secondary.

The fine-grained greenish rock has the same mineral assemblage, but the quantity of quartz varies somewhat. Such gneisses rich in epidote occur in other places also, and in some samples remnants of bluish green hornblende and biotite have been found together with the epidote.

Structurally these epidote-gneisses are distinguished from the normal gneisses by the recrystallization being more pronounced. Also on account of this it is difficult to conclude anything about the composition of the rocks before the metamorphism.

Leucocratic gneisses and mylonites. Some gneisses are distinguished by containing dark minerals to a very small extent only. They are found scattered throughout the field at various localities, but common to them all is that they occur in regions where the crushing has been especially strong. They are traversed by joints and cleavages where chlorite and epidote have been formed. The majority contain some ore and titanite, and muscovite may also occur, but no other dark minerals. The main constituents are plagioclase, potash feldspar, and quartz. The proportion between quartz and potash feldspar varies, as does the content of quartz.

Since these rocks are always attached to crushing zones, it is natural to regard the absence of dark minerals as a result of the crushing, and not as a primary characteristic.

A couple of types will further be described. A mangeritic mylonite-gneiss from Brekkenipo is a typical crushed rock, occurring as a shear zone in the anorthosite. It is an 'eyed' gneiss with reddish feldspar 'eyes', traversed by a lot of joints with chlorite and epidote. The chlorite is a penninite. Of other minerals only a little ilmenite with a marginal zone of titanite and leucoxene is found. Under the microscope the cataclastic 'eyed' structure is clearly seen. The ground mass is extensively broken up and fine-grained, and consists of quartz, plagioclase, and potash feldspar. The plagioclase is relatively fresh; a determination of the index of refraction shows

$1.537\ \alpha_{Na} < 1.540$, ɔ: c. 20 % An. In the ground mass rather large lenticular grains of microperthite and plagioclase are seen, also crushed and traversed by joints with chlorite. Chlorite and epidote also surround the lenses, and the 'eyed' structure is created in this way.

Also a sample from a shear zone in the anorthosite of Larsfonnfjellet has been examined. This, too, contains very little dark minerals. The 'eyed' structure is not so pronounced here, but under the microscope it shows clear signs of crushing. Large grains of acid plagioclase lie in a fine-grained ground mass. The plagioclase shows Karlsbad twins. The quantity of quartz has been estimated at more than 12 %. A little muscovite occurs. Also this rock is traversed by a lot of joints where chlorite and epidote have been formed.

At Vetlestøl just beyond the border to the mica schist there is a similar rock. Macroscopically it is green-coloured. This colour originates in the epidote and chlorite, which also here, are chiefly found on joints. The chief mineral is plagioclase, an acid oligoclase $\alpha_{Na} < 1.536$, ɔ: c. 15 % An. It is not quite fresh having inclusions of sericite. A little microperthite and quartz are also found. Of dark minerals it contains a little biotite, ore, and orthite-epidote. It has the usual crushed structure.

According to Waters and Campbell's scheme these rocks are mylonite-gneisses. They are cataclastic rocks where an extensive recrystallization has taken place, especially along the planes of schistosity. Also the true mylonites exhibit the same absence of dark minerals. The definition of mylonite used by Waters and Campbell has been suggested by Lapworth in 1885. It runs as follows: "Mylonites may be described as microscopic friction breccias with fluxion texture, in which the interstitial dusty, silicious and kaolinitic paste has only crystallized in part.« When the material has been so finely crushed it is difficult, even in the microscope, to separate the individual constituents, and the degree of recrystallization is difficult to determine.

The mylonites are distinguished from the mylonite-gneisses by not being schistose; they are, as a rule, hard and massive rocks. The majority have a reddish hue that often is so deep, that it suggests the colour of jasper. An exception is the mylonite at Ulvund, described above, which has a bluish tinge. Under the microscope some few porphyroclasts of feldspar in a finely crushed ground mass are seen. Of dark minerals the following have been determined:

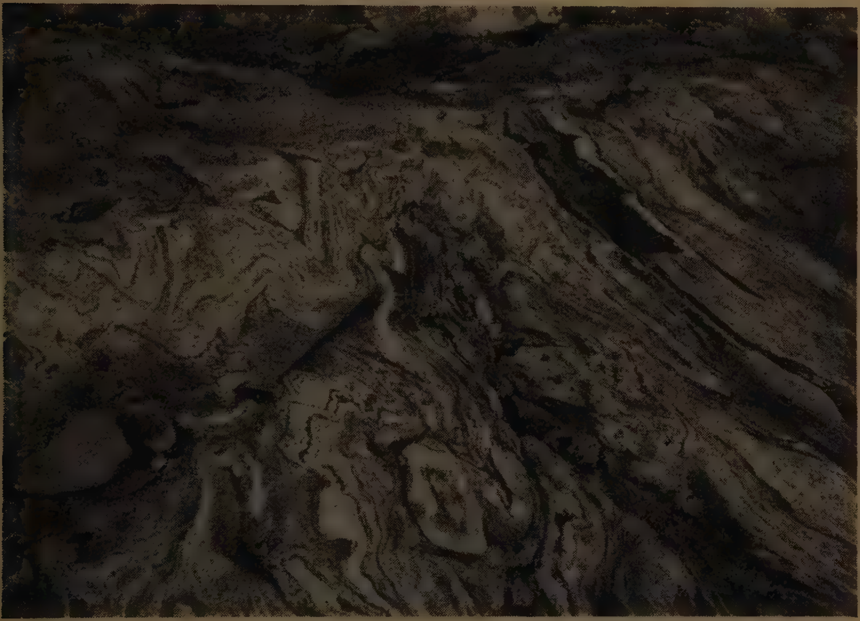


Fig. 24. Gneiss with flow folds, Fimbulfjell.

Chlorite and epidote; besides ore, titanite, and graphite, and in one single case remnants of hornblende. The dark minerals always constitute a very small part of the rock.

It accordingly is evident that the dark minerals disappear gradually by increasing crushing. As mentioned above there is no reason to suppose that these rocks originally have had another composition than the rest. The dark minerals, pyroxene, hornblende, or biotite must have been destroyed during the crushing. They must have been dissolved during the metamorphism, and chlorite and epidote have been formed later, in their places. These minerals are found, as mentioned above, chiefly on joints. In other words a leaching of the rocks has taken place, the Fe- and Mg-content being concentrated on joints, and the rocks otherwise lacking dark minerals. The mechanical deformation thus leads to a certain migration of the constituents in the rock, to a metamorphic differentiation.

Banded gneisses. Finally it will be mentioned that the gneisses in some regions much resemble the migmatites. Typical flow folds show that the rocks must have been plastic. They are banded, and light and dark bands alternate (Fig. 24).

A microscopical examination shows that the mineral assemblage is the same as in the other gneiss rocks, and therefore they may not be regarded as migmatites in the proper sense of the word. Structurally they are distinguished from regular gneisses, showing little or no crushing, but an extensive recrystallization. Such migmatite-like types have also been described from the mangerite gneisses of the Bergen field (Kolderup 1940).

FACIES OF THE MANGERITE-GNEISSES

The typical mineral assemblage of the gneisses is: Albite or acid oligoclase, potash feldspar, and quartz, and of dark minerals: Biotite, muscovite, chlorite, and epidote.

This corresponds to the epidote-amphibolite-facies.

A COMPARISON BETWEEN THE GNEISSES AND THE IGNEOUS ROCKS

The facies of the mangerites are, as mentioned above, hornblende-gabbro-facies. The chief minerals are: Uralite, basic oligoclase, potash feldspar, biotite, and ilmenite.

A comparison between the mineral assemblages of the non-metamorphic types and of the metamorphic types will show what reactions have taken place during the metamorphism.

It cannot be taken for granted that all the gneisses have been mangerites. The quantitative relation between the minerals has certainly varied, but the mineral assemblage has nevertheless been the same, so that purely qualitatively a good idea may be obtained in this way of what sort of reactions have taken place.

The transition from hornblende-gabbro-facies to epidote-amphibolite-facies is characteristic by the reaction:

Anorthite → clino-zoisite (or epidote).

The plagioclase of the mangerites has normally an An % of c. 22. In the gneisses 14—15 % is common, but the content of anorthite may sink to below 10 %. Hornblende is not found in the mangerite-gneisses, and it is therefore reasonable to suppose that Ca from it takes part in the reaction so that we get the following equation (Vogt 1927):

$$3CaAl_2Si_2O_8 + CaO + H_2O \rightarrow 2HCa_2Al_3Si_3O_{13}.$$

If CaO from the hornblende does not enter, quartz will be extricated. This reaction has taken place in the gabbro during the saussuritization of the plagioclase. Here the hornblende is preserved, and epidote and quartz have been formed of the anorthite component of the plagioclase. The quartz is seen here as small grains within the saussuritized ground mass. In the mangerite-gneisses the hornblende has disappeared and the plagioclase is not so seriously affected as in the amphibolites. Here, therefore, it is reasonable to suppose that the first reaction has taken place.

Biotite is also present in the non-metamorphic types, but it is here usually found as a coronal formation around the ilmenite. It marks the schistosity in the gneisses, and it is more abundant here. Consequently biotite must in this case have been new-formed at the expense of the other minerals, especially uralite. To this process a supply of potash is required. The only potash source in question here is the potash feldspar. Here a great deal of reactions must have taken place, of which Vogt (1927) gives some of the most important. Biotite may be formed at the expense of plagioclase, potash feldspar, and pyroxene. In all these reactions quartz will be formed as well. The reactions must at any rate partly explain why the gneisses contain more quartz than the fresh mangerites.

The potash feldspar of the non-metamorphic types is predominantly a microperthite. In the gneisses microcline and microcline-microperthite prevail. The course of the development may vary here.

The spindles of the fresh microperthites are lengthened and smooth. In the gneisses it may be seen in some cases that the spindles grow at the expense of the potash feldspar. These enlarged spindles are usually irregular. Cases of double perthite also occur. New spindles are formed by the metamorphism instead of the old ones being enlarged. Also these new-formed spindles are always irregular, clearly formed by replacement. The plagioclase of these new-formed spindles is acid; as a rule it is an albite, and often it is filled with sericite and epidote. Contemporaneously with the growth of the spindles we usually get an incipient microclinization. Most frequently the first microcline lattice is formed at the edge of the grain, and little by little the entire grain is transformed to microcline. Microcline-microperthite often occurs as a transitional link.

In the gneisses microperthite grains with very fine spindles are also found, often in rocks where the mechanical metamorphism has prevailed. The original spindles appear to have been pressed together. Also these types generally pass into microcline by increasing metamorphism.

Besides the albitization, mentioned above, of the potash feldspar by the growth of the spindles, an albitization from the edges is also common. A zone of albite is formed around the microcline grain, and from this zone tongues of albite penetrate into the grain.

The normal development, then, is that the microperthite is transformed into microcline, and that the latter is albitized. Hereby potash will be released, which will be spent in the formation of the biotite, or muscovite if little Fe and Mg is present.

The ilmenite of the mangerites usually has a corona of biotite, while in the gneisses it is surrounded by titanite and leucoxene. By increasing metamorphism the ilmenite is reduced while the titanite grows. The titanite must have been formed at the expense of the ilmenite by a supply of lime and silica.

These are in outline the changes of the mineral assemblage, occurred during the metamorphism. Whether changes in the chemical composition have taken place is difficult to state, so long as analyses of the rocks are not available. The most prominent feature of the gneisses is that they contain more quartz and more biotite than the non-metamorphic types. The question then is whether the quantity of potash and silica released by the process mentioned, is great enough to explain this difference. Without chemical analyses this is impossible to say with certainty, but judging from the microscopical examination I am inclined to believe that there must have been a supply of silica and potash during the metamorphism. It has been suggested above that these materials may have been supplied from the depth, that accordingly an incipient granitization of the rocks may have taken place during the metamorphism. This is, as far as I can see, the most reasonable explanation.

STRATIGRAPHY AND TECTONICS

The main features of the stratigraphy and tectonics of the area have been mentioned above, but here a general survey of them will be given. The stratigraphy is very simple. Undermost lies Cambro-Silurian mica schist, allochthonous above this lies a massif of the rocks of the anorthosite kindred. Thrust into mica schist lie also several areas of quartz schists. Between Skreia and Mykjedalsvatn the mangerite-gneisses border against these rocks.

Above has been mentioned that a distinction has been made between primary and secondary structures. The structures around the anorthosite massif are interpreted as primary in the sense that they have been formed during the *mise en place* of the rocks. These structures have been mentioned above, and will not be further described here. See also the structural map of the area surrounding the anorthosite massif as well (Fig. 4).

The directions of the flow lines, shear zones and folding-axes are brought out by the map. Quite a lot of the flow lines have NW—SE direction, that is the same direction as the stretching of the gneisses. The directions, however, vary so much that conclusions from this circumstance cannot be justified. Also the folding-axes have partly NW—SE direction. It has been mentioned above that they show a tendency to be normal to the boundaries, but nor here anything can be said with certainty. This applies to the shear zones as well. It might be tempting to see a connection between this NW—SE direction and the direction of the gneisses, and consequently with the thrusting, but because of the great variations of the measurements, I do not think it permissible to draw any conclusions from them.

The schistosity and the flow lines of the gneisses are regarded as secondary, metamorphic structures, as well as their crushing zones and those of the anorthosite area. These structures have been formed during the thrusting to which it stands in relation.

The strike of the gneisses is NW—SE and the dip towards NE. This direction deviates from the common Caledonian NE—SW direction. This may be connected with the area being situated at the edge of the great thrust masses, where such irregularities will more easily arise. The strike of the mica schists is on the average N—S, while the normal NE—SW direction is found in the quartzites and

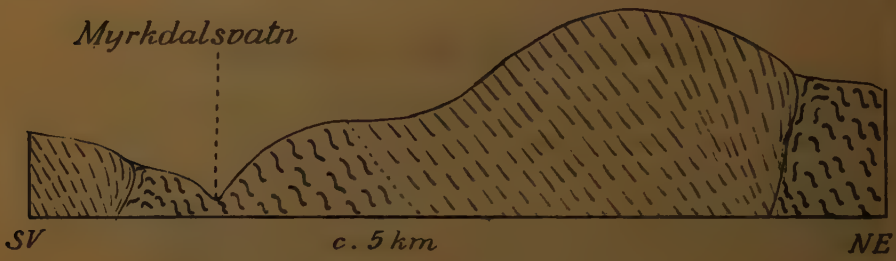


Fig. 25. Section showing the bending at the thrust front.

quartzitic schists situated farther west. The dip is everywhere NE. The lowest levels accordingly are found in SW, and the highest in NE.

Thrust direction, stretching, and folding-axes are sub-parallel and run NW—SE, their direction being usually 10°—20° more westerly than the strike direction. This thrust direction is common in the large overthrustings of the Caledonian Mountain Range in Scandinavia (Th. Vogt 1927, Strand 1938, Kvale 1941). That a thrusting really has taken place is obvious. The schistosity bends at the thrust front and runs parallel to the front in the border (v. map and profile section fig. 25). Where the thrust plane crops out there is as a rule a crushing of the rocks on both sides, but most pronouncedly at the thrust front. Kvale (1941) defines thrust plane and thrust front in the following way: "Thrust plane is a boundary plane of a thrust mass (usually no level plane). Thrust front is the foremost border of a thrust mass, that is to say not any line of intersection between a thrust plane and the surface of the earth." In these senses the words will be used here also. Measurements of gliding striæ show that the thrusting has taken place from NW towards SE.

The stretching expresses the maximum prolongation caused by the metamorphism. It is here parallel to the thrust direction.

The schistosity is marked by the new-formation of mica, as a rule biotite. The biotite frequently also marks the stretching.

The lines of intersection between the earth's surface and the thrust plane or planes (there are usually more than one) appear clearly in the ground. A clear thrust front runs along the border against the mica schist from the west side of Uppheimsdalen to Mykje-

Fig. 26. Thrust front at Skreia.

dalsvatn and farther southwards past Syndve. Clear thrust fronts are also found north of Skreiavatn and at Kasetseter. The thrust fronts run approximately vertically to the thrust direction, and there is a clear crushing and a heaping of rocks near by (fig. 26).

Lines of intersection between the topographic surface and the thrust plane appear also in all the places where the gneisses border aganist the mica schist. They are as a rule clearly marked in the ground. The harder and more resistant gneisses rest on a mica schist that is softer and weathers more easily: This often results in a landslip, and the typical picture is steep precipices with a talus below (fig. 27).

A further examination of these precipices shows that there is here a typical transitional zone between the mica schist and the gneisses. Such a zone will be described from the west side of Finnbufjell. The gneiss is here a regular 'eyed' gneiss with reddish feldspar 'eyes'. Nearer to the thrust plane the gneiss becomes more fine-grained, and the 'eyes' become fewer. Finally they disappear entirely, and the rock becomes a fine-grained, slaty gneiss. The

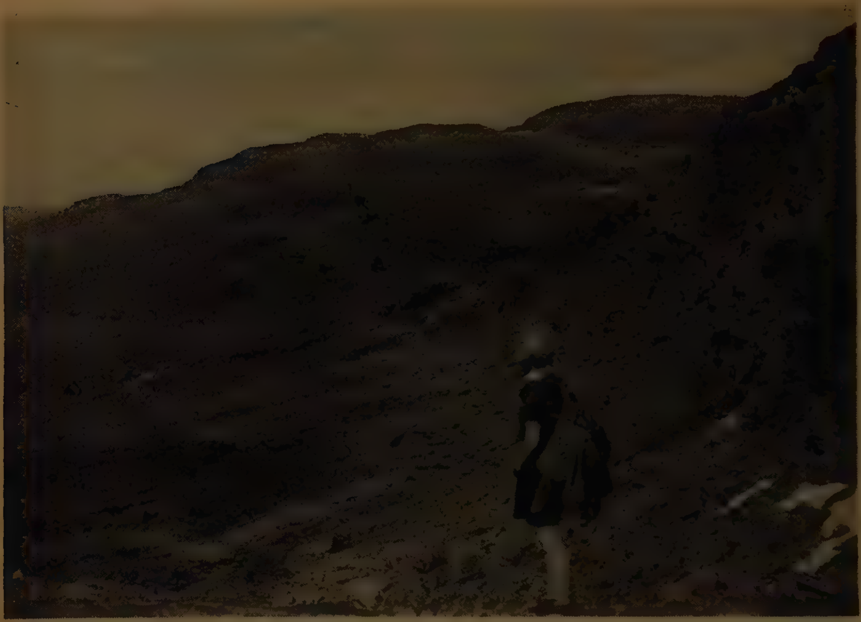


Fig. 27. Line of intersection between the surface and the thrust plane.

zone from the 'eyed' gneiss to the fine-grained gneiss has here a vertical extent of c. 10 m. Under the slaty rock lies a layer of mica schist. The border between them is quite sharp. Then comes a zone of fine-grained, slaty gneiss, therupon a zone of mica schist, still separated by sharp borders. There is a total of three zones of slaty gneiss, their vertical extent varying from 0.5 to 1 m. Under the microscope these rocks prove to be extensively crushed mangeritic gneisses, and the crushing is so thorough that they approach mylonites.

A thin section of the border between the mica schist and the slaty mangerite-gneisses has been made. Also under the microscope the border is entirely sharp. It is marked by the sudden increase of the content of mica in the mica schist. Whether a hybridization has taken place is not easy to decide with certainty, as the mangerite-gneisses contain in advance both quartz and mica. On the other hand the mica schist may contain a little albite. Albite is found in all the samples of mica schist from the boundary areas, but it is also found in several samples outside. These boundary rocks are everywhere so fine-grained that it is very difficult to determine the quantitative relations. Above has been demonstrated that the T- and P-conditions here have been high enough to make a considerable recrystallization possible. Accordingly also a hybridization at the boundary may possibly have taken place.

The conditions described here from Finnbufjell are typical, and are found wherever the thrust plane crops out. They relate a good deal about how the thrusting must have taken place.

We must imagine that the gneisses have entered along the planes of schistosity of the mica schist and glided along these planes. The mica schist has served, as it were, as a lubricant during the movement. In the vicinity of the thrust plane the gneiss has become slaty and fine-grained. It cleaves well and forms flagstones. This corresponds perfectly to the conditions found in the slate quarries of the Voss district (described by N. H. Kolderup 1933). The slates are found here in the boundary region between the mica schist and the over-thrust rocks. The pronounced schistosity of the boundary areas is explained as a result of the pressure during the thrusting. In Vossestrand flagstones or slates are not found in such quantities that the occurrence is workable, but in other respects the conditions are perfectly analogous.

Layers of mica schist and gneiss in alternating succession are also found here, which only shows that the thrusting has taken place in several levels. A peculiar point is that the borders between the zones of mica schist and gneiss always are quite sharp. This seems to suggest that the over-thrust rocks must have met with slight resistance during the thrusting. The gneiss zones have a linear course, and neither foldings nor faultings have been observed along them. This, too, points in the same direction.

Quite different are the conditions at the thrust fronts. There always is an extensive brecciation along the typical thrust fronts, and we find breccias and mylonites. The crushing predominates both in the gneisses and in the mica schist below, though the border between these rocks is quite sharp. Fine exposures of the boundary are found at Ulvundsøyni, by Svartselv in Uppheimsdalen and north of Skreiavatn (Fig. 28).

A typical feature of the thrust front is that the schistosity always follows the boundary, and that the dip changes. Usually the dip is towards the east, but along the thrust fronts the dip is westerly, both

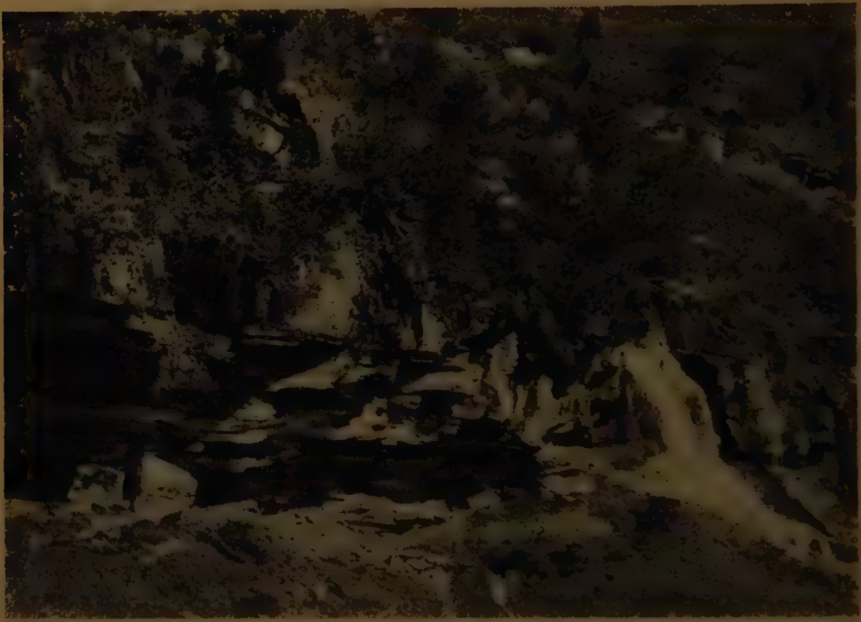


Fig. 28. Border between mica-schist and gneiss. N. Skreia.

in the gneiss and in the mica schist. A heaping of rocks has taken place along the fronts. Along the other outcrops of the thrust plane there is no such heaping, and the schistosity of the gneisses usually will cross the boundary also here. This is only what might have been expected when it is borne in mind that the westerly limitation of the over-thrust rocks represents a quite casual intersection-line between the thrust plane and surface of the earth, while the thrust fronts represent a real front line where the movement of the thrust mass, for some reason, has come to a halt.

The direction of the thrust fronts always is approximately vertical to the thrust direction. A number of crushing zones in the interior of the massif show a great similarity to the thrust fronts. They too run, as a rule, vertically to the thrust direction, and may in some way be regarded as sort of secondary thrust fronts. They are most common in and around the anorthosite massif, but are also found in other places of the field. The greatest of these zones are marked on the map.

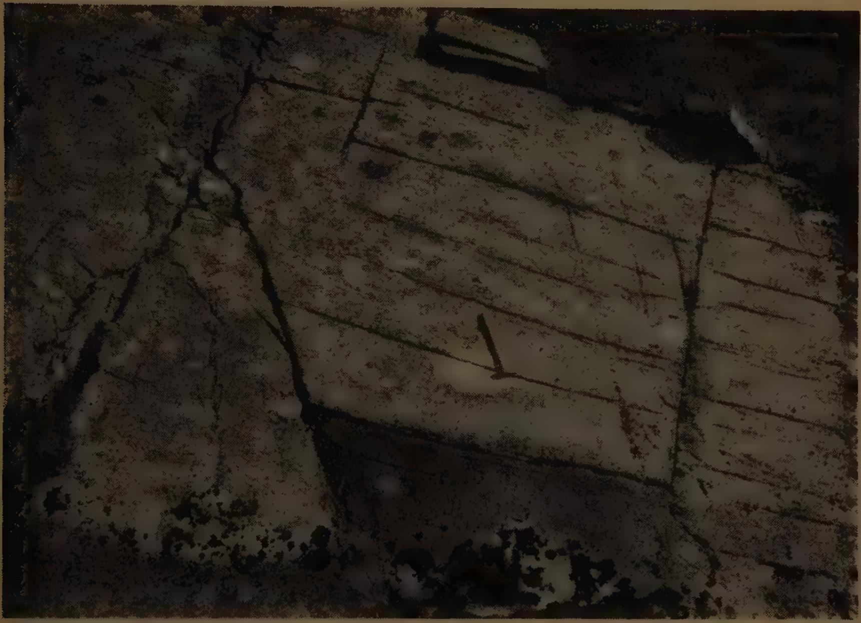


Fig. 29. Directions of joints in gneiss, Solliane.

A thrusting is a movement along one or more gliding planes. According to the dip of the gliding plane the movement may run upwards or downwards. There is much in favour of the view that we have to do with a downward movement here. As mentioned above, there is a pronounced stretching in the gneiss rocks, and this stretching is parallel to the thrust direction. Such a direction of the stretching may be more easily understood if the movement rather has had the character of a gliding.

The area is situated on the north-west side of the Sogn—Jotunheimen syncline, and the direction of the movement is towards the syncline. It is not the least unreasonable to imagine that the thrust plane has inclined towards the syncline during the movement. The force of gravitation must then have been a contributory factor of the movement, and we may imagine that the rocks have glided down into the syncline. Accordingly we may easily understand that they will be drawn out in the direction of the movement. The heaping of rocks along the thrust fronts is also explained in a natural manner by this assumption.

Joints. In the gneiss area there are several pronounced joint systems of which cross joints and strike joints are most common.

The valleys practically always follow one of these directions, most frequently the strike direction. This appears clearly on a topographical map. As mentioned above the strike direction of the quartz and the quartzitic schists to the west is NE—SW, while the main strike direction of the rocks of the anorthosite kindred is NW—SE. This fact is distinctly reflected in the formation of the valleys, which change their direction according to the strike. Examples of strike valleys are Brekkedalen, Kvassdalen, and the valley at Myrkedalsvatn, while Myrkedalen itself is a cross valley.

The direction of the joints may be recognized on a larger or smaller scale. Strike and cross joints may be so strongly developed that the rock is apt to split up. Diagonal fissures may occur too, though they are not so common (Fig. 29).

Minerals of the joints. Chlorite is most common without comparison, and it is found on strike and cross joints as well as on thrust planes. Epidote is less common. By the crossroads at Tveite an orthorhombic zeolite-mineral has been found as joint filling. A measurement of the indices of refraction gave the following figures:

$\gamma_{Na} \sim 1.500$

$\alpha_{Na} \sim 1.487$

$\gamma - \alpha \;\; 0.013$ This corresponds to a natrolite.

CONDITIONS OF PRESSURE AND TEMPERATURE DURING THE METAMORPHISM

In the exposition above a distinction has been made between mechanical and chemical metamorphism or between brecciation and recrystallization. Which of these processes has played the greater part is not easy to state, but it is quite certain that an extensive recrystallization has taken place. Mylonite-gneisses is the most adequate term for the majority of the gneiss rocks. This shows that pressure and temperature during the consolidation cannot have been inconsiderable. On the other hand the friction during the movement will increase the temperature considerably, so that no great depth is needed to bring about a recrystallization. Several facts seem to suggest that the thrusting has taken place at a high level. In the first instance the crushing zones with breccias and mylonites along the thrust fronts must be mentioned. If the thrusting had taken place at a greater depth, we would probably get a folding here with axes parallel to the front. Such conditions have been described by Kvale (1941) from the quadrangle Bergsdalen. Here we have got a crushing instead. The rock must have possessed, however, a certain degree of plasticity. Where the schistosity is preserved it is conformable to the front, and the dip changes at the boundary. A bending of the direction of the rocks has taken place at the fronts. The further development of such a bending to a folding with axes parallel to the front is not considerable.

As mentioned above, the gneiss area also contains rocks that have been plastic during the deformation (Fig. 24).

It appears, then, that the degree of metamorphism varies greatly. There are all transitions from fresh igneous rocks over the various gneiss rocks to mylonites and breccias on one hand and purely plastically deformed rocks on the other.

This may be due partly to primary differences in the rocks. This is, however, not sufficient to explain the great variations. It must also be presupposed that the metamorphism has not been equally strong in the various localities.

This is just what might have been expected when the metamorphism is regarded as a result of the thrusting.

The result of a thrusting seldom is a homogeneous distribution of the metamorphism in the rock. Usually definite zones are found in a thrust mass where the metamorphism has been especially strong, while the rock may be altered only a little between these zones.

Besides along the thrust fronts the crushing zones are first of all found in the vicinity of the anorthosite area. The anorthosite is at the same time far the least altered of the rocks. The anorthosite has reacted essentially in another manner to the metamorphism than the intermediate and acid rocks. During the thrusting the anorthosite massif has offered resistance like a solid block, and has chiefly been subjected to mechanical deformation. This difference of degree of metamorphism is not due to the fact that the rocks have been at various levels during the thrusting. Intermediate and acid rocks on the same level and on higher ones than the anorthosite have become gneissic. It must then be due to different capabilities of resisting the effects of the metamorphism.

The new-formed minerals may give an idea of what temperatures prevailed during the metamorphism. It is first of all quartz and biotite that have been new-formed besides some muscovite. According to Eskola (1939) muscovite (sericite) is formed in an interval of temperature from 500° to 225°, possibly at somewhat lower temperatures, too.

The new-formation of biotite requires a somewhat higher temperature. A definite interval of temperature cannot be indicated here, *i. a.* because the composition of the biotite varies rather much.

It may be taken for granted, at any rate, that a series of rocks of several hundred meters thickness have lain above the present surface. Holtedahl (1944) characterizes these conditions in the following way: "The depth and temperature conditions of the moving igneous masses were no doubt closely connected with the primary thickness of the sedimentary series. The Hardanger—Jotunheim area, the Archean basement-complex of which was not at all or only to a very small degree petrographically altered and mobilized during the main deformation processes, must be supposed to have had a much thinner cover of sediments than the districts to the NE — —." Further: "These masses which in the Jotunheim—Sogn district principally are made up of rocks belonging to the Bergen—Jotun-group of Goldschmidt, with anorthosite as a typical constituent, farther south of largely granitic rocks of various types, have been moved in a more or less solidified, cooled condition, . . . In the districts where they now occur, they moved along at moderate depth, . . ."

The conditions here are perfectly in accordance with Holtedahl's exposition.

THE AGE OF THE ROCKS OF THE ANORTHOSITE KINDRED

The boundaries between the mica schist and the rocks of the anorthosite kindred are everywhere secondary. Nothing can therefore be concluded from the boundary conditions as to the age relation between the two formations.

The only facts available for determining the age are the structural conditions within the over-thrust rocks. Above a distinction has been made between primary and secondary structures. It must be regarded as certain that the secondary structures are Caledonian. I think I have demonstrated that the structures around the anorthosite massif must be older than the Caledonian structures of the gneisses. Consequently it may be concluded that the rocks of the anorthosite kindred must have been formed in the pre-Caledonian or early Caledonian period. The extensive brecciation shows that the masses must have been more or less solidified during the thrusting. A certain period of time must have separated the intrusion of the magma and the thrusting of he rocks. However, no data are available that might suggest the duration of this period.

SAMMENDRAG

Dette arbeide omhandler bergarter av anorthositstammen på kartblad Vossestrand, C 32 V. Feltet ligger på NV-siden av Sogn—Jotunheimsynklinalen og i den vestligste utkant av de store skyvedekker. Beliggenheten fremgår av kartskissen (fig. 1). Det er også gitt en kort beskrivelse av omgivende glimmerskifre, kvartsitter og kvartsskifre. Grensen mot glimmerskifrene er sekundære, det er vist at anorthositstammens bergarter er skjøvet på plass i sin nåværende stilling.

Det er funnet representanter for de fleste ledd av anorthositstammens bergarter. Utbredelsen av de forskjellige typer fremgår av kartet. Ved klassifikasjonen av bergartene er Nigglis system av 1931 lagt til grunn (se fig. 3). Bergartene av anorthositstammen er plasert i dette systemet der hvor de definisjonsmessig hører hjemme. Det er gjengitt definisjoner på de viktigste ledd innen stammen (se s. 137). For bergarter av syenodioritfamlien er navnet jotunit foreslått i stedet for V. M. Goldschmidts »jotun-norit«.

Området har vært med i den kaledonske fjellkjedefoldning, og bergartene bærer da også tydelig preg av dette. Bare mindre partier av friske eruptiver er bevart. Av bergarter som har motstått metamorfosens virkninger må først og fremst nevnes anorthositmassivet mellom Uppheimdalen og Brekkedalen. I umiddelbar nærhet av dette finnes det og en del mindre partier av intermediære og sure bergarter hvor mineralene er friske og de opprinnelige strukturer er bevart. Men for øvrig har metamorfosen vært gjennomgripende, de basiske bergarter er blitt omvandlet til amfiboliter og saussuritgabbroskifre

og de intermediære og sure bergarter til skifrige biotitgneiser. Disse siste er mest utbredt. Myloniter og breksier forekommer og.

Beskrivelsen av de friske eruptiver er gitt en relativt bred plass. Selv om de ikke er meget utbredt, er det i første rekke dem som gir nøkkelen til forståelsen av bergartenes genesis.

Det er utført 16 volumetriske analyser av eruptivene, med hensyn til disse og til bergartsbeskrivelsen for øvrig henvises til teksten s. 156—192.

Forholdene i og omkring anorthositmassivet mellom Uppheimsdalen og Brekkedalen er særlig interessante, og vil bli omtalt nærmere. Det er også laget et strukturkart over dette området (fig. 4). Anorthositen hviler i en skål av intermediære og sure bergarter, vesentlig mangeriter. Det er ingen skarp grense, men en overgangssone mellom anorthositen og mangeritene. Som overgangsledd opptrer dioriter, gabbroer og anorthositgabbroer. Denne overgangssonen er skifrig, og skifrigheten følger grensen. Strukturene her har en helt annen retning enn de kaledonske strukturer i gneisene. Dette viser at anorthositen må ha sunket i forhold til de omgivende mangeritiske bergarter. På NV- og SØ-siden av massivet er de to strukturer adskilt av en mylonitsone, Ø og V for massivet har strukturene samme retning. Det er ikke funnet gjennomsettende strukturer.

Inne i anorthositmassivet er det et større parti av en bergart som er blitt kalt gabbro »*in statu nascendi*«. Liknende bergarter er beskrevet av Balk (1931) fra Adirondacks, og betegnelsen er innført av ham. Mindre partier av gabbro og basiske utskilninger av pyroksen, hornblende og granat opptrer rikelig, men anorthosit er den dominerende bergart. Den kan være massiv med retningsløs kornet struktur, men svært ofte har den linjestruktur. Denne er da markert av små spindler av pyroksen, hornblende og granat. Det opptrer også skifrige sure og mangeritiske soner inne i anorthositmassivet, soner av skifrig anorthosit forekommer og. Disse sonene svarer til det som Balk har kalt »shear-zones«. Ganger og bånd av kvarts er meget utbredt, de opptrer først og fremst i grenseområdene. De er skifrige, og strukturene er konforme med strukturene i anorthositområdet for øvrig. Både feltiakttakelser og mikroskopiske undersøkelser tyder på at det her dreier seg om »magmatisk« kvarts, og derfor er betegnelsen silexit brukt om disse bergarter.

De opptredende eruptiver oppfattes som co-magmatiske. Her skal gjengis de viktigste trekk som taler for dette. At de intermediære og sure bergarter utenfor anorthositområdet er genetisk sammenhengende må ansees for helt sikkert. Feltiakttakelser viser at de forskjellige typer er intimt forbundet slik at det ikke lar seg gjøre å sette grenser mellom dem, de lar seg i regelen heller ikke skille fra hinannen makroskopisk. Undersøkelser i mikroskopet viser at de forskjellige typer er forbundet ved overganger, og at mineralselskapet stort sett er det samme med den basiske mikroperthit som karakteristisk mineral. Ser vi på den annen side på bergartene innen anorthositområdet, så er sammenhengen mellom dem og klar. Det er alle overganger mellom anorthosit, anorthositgabbro, gabbro og de basiske utskilninger. Mineralselskapet er og det samme, det er bare mengdeforholdet mellom mineralene som varierer. Spørsmålet er så om det er noen genetisk forbindelse mellom anorthositen og de mangeritiske bergarter. Det er en rekke trekk som taler for at dette er tilfelle. Det opptrer aldri skarpe grenser, men overgangsssoner mellom anorthositmassivet og de omgivende bergarter. Videre finnes det mangeritiske bergarter inne i anorthositen (shear-zones). Det omtalte anorthositmassiv ligger helt omgitt av mangeritiske bergarter og må betraktes som en »outlier« i forhold til det store anorthositmassiv ved Voss—Sogn. Dette forhold viser entydig at anorthosit og mangerit må være noenlunde samtidig dannet. Forekomsten av gabbro »*in statu nascendi*« er og en ting som taler for magmatisk differensiasjon. Det er en båndet bergart hvor lyse og mørke bånd veksler. I disse båndene finnes representanter for praktisk talt alle bergarter innen anorthositstammen. En betraktning over hvordan An % i plagioklasen varierer i de forskjellige bergarter peker og i samme retning. I de friske anorthositer har plagioklasen i regelen en An % på 60—63, mens den i mangeritene ligger på 20—23. Nå vet vi at i en plagioklassmelte er krystaller med An_{65} i likevekt med en smelte av sammensetning An_{25}. Dette er jo en meget pen overensstemmelse.

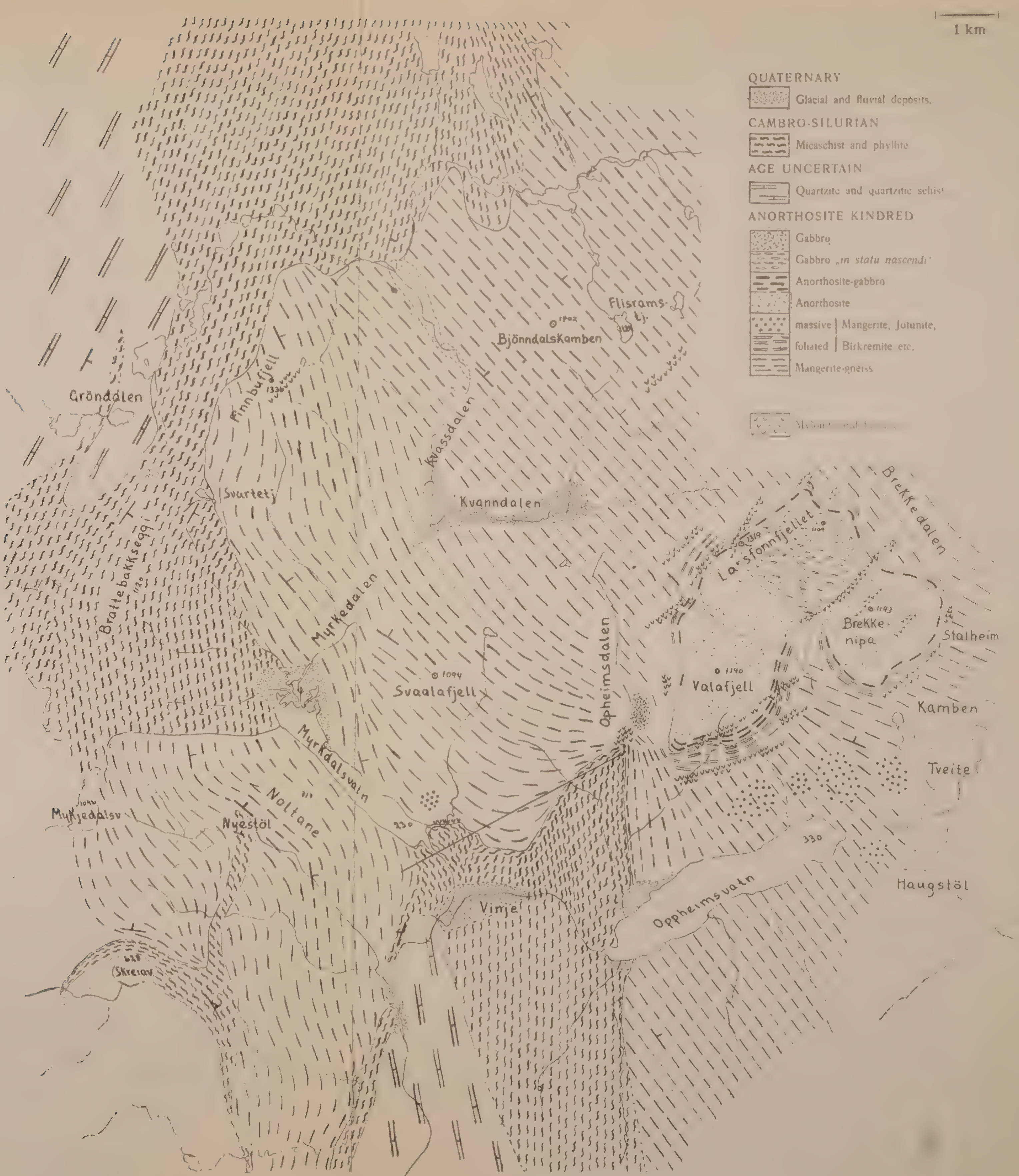
Av det som her er sagt skulle det fremgå at bergartene må være dannet ved magmatisk differensiasjon. Som nevnt viser bergartene stor likhet med dem i Adirondacks, og Balks teorier for den magmatiske differensiasjon der kan i stor utstrekning anvendes på anorthositstammens bergarter på Vossestrand. Med Balk må vi anta at anorthositen og anorthositgabbroen er skilt ut som krystalldifferensiater mens mangeritene er størknet fra et magma. Den sterke oppknusning i anorthositen (fig. 8) viser at det har foregått bevegelser i »krystallgrøten«. »Shear-zonene« oppfattes som rester av modermagmaet (the mother liquor) som er blitt presset ut. Silexitene oppfattes endelig som siste sure ledd av erupsjonen.

Den skifrige grensesone rundt anorthositmassivet viser mange interessante trekk. Som nevnt danner den samtidig en overgangssone mellom mangeritene og anorthositen. Begynner vi underst i mangeritene er det fra dem en gradvis overgang til dioritiske bergarter, og derfra videre til gabbro og anorthositgabbro. I denne sonen har det vært en utstrakt rekrystallisasjon, det er nydannet hornblende og biotit. Også plagioklasen er i stor utstrekning rekrystallisert, den er surere enn plagioklasen i det indre området, og samtidig har bergartene mere epidot. Det er her en tydelig sammenheng mellom An % i plagioklasen og epidotmengden i bergarten. Epidoten forekommer i store korn som ofte er idiomorft begrenset. Denne grensesonen oppfattes her som en reaksjonssone mellom anorthositen og de omgivende mangeriter dannet idet anorthositen sank i forhold til restmagmaet (the mother liquor). Den representerer også en sone hvor temperaturen må ha vært lavere enn i det indre, forholdene svarer altså til dem en finner ved en vanlig pluton med de høyeste temperaturer i det indre, og lavere temperaturer ut mot grensen. Bergartene i det indre av anorthositområdet er i granulitfasies, mens bergartene i grenseområdene er nærmere hornblende-gabbrofasies (fig. 17). For de intermediære og sure bergarter utenfor anorthositområdet gjelder det at enkelte av de friskeste typene er i granulitfasies, men stort sett står disse bergarter nærmere hornblende-gabbro-fasies. Dette tyder på at t- og P-forholdene under »størkningen« har beveget seg fra granulitfasies mot hornblende-gabbro-fasies. Denne utviklingen er pent illustrert ved en rekke randsoner som oppfattes som senmagmatiske reaksjoner. Uralitisering av pyroksen er særlig pent utviklet (fig. 14).

Metamorfosen. Som nevnt er det skifrige biotitgneiser som utgjør størsteparten av berggrunnen på kartbladet. Sammenhengen mellom disse gneisene og de beskrevne friske eruptivene er klar. Metamorfosen har ikke vært like gjennomgripende overalt, og en kan følge de forskjellige stadier fra friske eruptiver i hornblende-gabbro-fasies til skifrige biotitgneiser i epidot-amfibolit-fasies. Metamorfosen sees som resultat av en skyvning, en kan stort sett skille mellom en mekanisk og en kjemisk virkning av den. Bergartene har vært utsatt for en sterk oppknusning, i enkelte tilfelle er det dannet myloniter og breksier, men kataklastiske øyengneiser med øyne av kalifeltspat eller plagioklas er mest utbredt. Med stigende metamorfose avtar størrelsen av øynene, og i de laveste nivåer lengst i SV er det overveiende finkornete, planskifrige gneiser uten øyne. Bildene på fig. 19—23 illustrerer utviklingen fra relativt friske eruptiver over øyengneiser til mylonit. Blant gneisene forekommer og båndete migmatitliknende bergarter som tydelig er blitt plastisk deformert (fig. 24). De viktigste forandringer i mineralselskapet under metamorfosen er at gneisene har betraktelig mere kvarts enn de friske eruptiver og at biotit er det dominerende mørke mineral. De friske eruptiver har pyroksen og hornblende. Det foreligger ikke kjemiske analyser av disse bergartene, og det er derfor vanskelig å avgjøre om det har vært noen stofftilførsel under metamorfosen. Men ved å dømme ut fra den mikroskopiske undersøkelse er jeg tilbøyelig til å tro at det må ha vært en tilførsel av kali og kiselsyre. Som nevnt er bergartene i anorthositmassivet lite påvirket av metamorfosen, men virkningene av den kan spores her og. Bergartene er overalt sterkt oppknust, men det er ofte vanskelig å avgjøre hva som er proto- og hva som er kataklase. At det har foregått bevegelser etter størkningen er sikkert, det viser bl. a. en rekke forkastninger i de de mørke bånd og i »shear-sonene«. Dessuten er oppknusingssoner meget utbredt i og omkring anorthositmassivet. De viktigste av disse sonene er avsatt på kartet, de har en tydelig relasjon til skyveretningen. At det her virkelig har foregått en skyvning, framgår tydelig. Ved skyvefrontene bøyer planstrukturene og går parallelt med fronten i grensen (fig. 25). Hvor skyveplanet kommer fram i dagen, er det i regelen en sterk oppknusning på begge sider av det, mest utpreget er denne oppknusningen ved skyvefrontene. En rekke målinger av glidestriper tyder på at skyvningen har foregått fra NV mot SØ, denne skyveretning er vanlig ved de store overskyvninger i den kaledoniske fjellkjede i Skandinavia. Strekning og foldningsakser går også NV—SØ, deres retning er gjerne 10°—20° mere vestlig enn strøkretningen. Skjæringslinjene mellom jordoverflaten og skyveplanet er i regelen godt markert i terrenget. Det typiske bilde er bratte hamrer med nedraset materiale nedenfor (fig. 27). Et nærmere studium av disse hamrene viser at det er en overgangssone mellom gneisene og den underliggende glimmerskifer. Lag av glimmerskifer veksler med lag av en finkornet, planskifrig bergart som under mikroskopet viser seg å være en sterkt oppknust mangeritgneis. Fra denne planskifrige gneisen er det en gradvis overgang til vanlig øyengneis. Dette viser at skyvningen må ha foregått i flere plan. Grensene mellom sonene av glimmerskifer og gneis er alltid helt skarpe, og de har ett rettlinjet forløp, det er ikke iakttatt foldninger eller forkastninger langs dem. Dette tyder på at de overskjøvne bergarter må ha møtt liten motstand under bevegelsen. Vi må tenke oss at gneisene er trengt inn langs skifrighetsflatene i glimmerskiferen og glidd langs disse. Glimmerskiferen har tjent som smurning under bevegelsen, i nærheten av skyveplanet er gneisene blitt planskifrige og finkornete på grunn av presset under skyvningen. Retningen av skyvefrontene er alltid tilnærmet loddrett skyveretningen, samme retning har også de oppknuste sonene inne i massivet. Disse kan på en måte betraktes som en slags sekundære skyvefronter. En skyvning er en bevegelse langs en eller flere glideflater, alt etter som fallet på glideflaten er kan bevegelsen bli oppover eller nedover. Det er meget som tyder på at vi her har hatt en bevegelse nedover. Området ligger jo på NV-siden av Sogn—Jotunheimsynklinalen, og bevegelsesretningen er inn mot denne. Det er ikke noe urimelig i å tenke seg at skyveplanet har hellet ned mot synklinalen på den tid bevegelsen foregikk. Tyngdekraften vil da ha vært en medvirkende faktor under bevegelsen, og denne vil like meget ha hatt karakteren av en glidning. Strekningen i bevegelsesretningen og oppstuvingen av materialet ved skyvefrontene får sin naturlige forklaring ut fra denne teori.

Det er en viss regional fordeling av metamorfosen idet vi finner de mest omvandlete bergarter i de laveste nivåer lengst i SV. Men det er mange unntagelser fra denne regel. Det vanlige er at en finner bestemte soner hvor oppknusningen har vært særlig sterk, mens bergartene imellom dem kan være lite omvandlet. Dette er rimelig når metamorfosen sees som et resultat av skyvningen. Nå er det påfallende at oppknusningssonene foruten ved skyvefrontene først og fremst finnes i nærheten av anorthositmassivet. Samtidig er anorthositen uten sammenlikning den minst omvandlete av bergartene. Under skyvningen har anorthositmassivet stått imot som en fast blokk, bare enkelte soner er blitt mekanisk deformert. Det er og verdt å legge merke til at de friske intermediære og sure eruptiver som forekommer ligger i umiddelbar nærhet av anorthositmassivet. I bergartene utenfor anorthositmassivet har det ved siden av oppknusningen vært en utstrakt rekrystallisasjon. Det er først og fremst kvarts og de mørke mineraler som har lett for å rekrystallisere mens feltspat under samme forhold blir oppknust. Mylonitgneiser er den

GEOLOGICAL MAP OF THE VOSSESTRAND DISTRICT



Johanne Hødal: Rocks of the Anorthosite kindred in Vossestrand.

betegnelse som passer best for de fleste av gneisbergartene (se s. 207). Dette viser at t- og P-forholdene under skyvningen ikke kan ha vært ubetydelige. På den annen side vil friksjonen under bevegelsen øke temperaturen betraktelig, så det kreves ikke så stort dyp for å få rekrystallisasjon. Men en må i hvert fall kunne gå ut fra at det har ligget en bergartserie på flere hundre meter over det som i dag er overflaten.

Det er vist at strukturene i anorthositmassivet må være eldre enn de kaledonske strukturer i gneisene. Av dette følger da at anorthositstammens bergarter må være av pre-kaledonsk eller tidlig kaledonsk alder. Noen andre holdepunkter for en aldersbestemmelse er ikke funnet.

Bergens Museum, oktober 1944.

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Abbreviations:

N. G. T. — Norsk Geologisk Tidsskrift.
N. G. U. — Norges Geologiske Undersøkelse.
V. S. S. — Videnskapsselskapets skrifter (Videnskaps-Akademiet, Oslo).
B. M. Å. — Bergens Museums Årbok.
B. M. S. — Bergens Museums skrifter.
T. M. P. M. — Tschermaks mineralogische und petrographische Mitteilungen.
S. M. P. M. — Schweizerische mineralogische und petrographische Mitteilungen.
G. F. F. — Geologiska Föreningens i Stockholm Förhandlingar.

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OM LEPTITBEGREPET

AV

CHR. GLEDITSCH

Ordet leptit er avledet av det greske ord λεπτός, som betyr fin. Ordet ble innført i petrografien av den svenske geolog Hummel i 1875 (1). Han foreslo leptit som fellesbetegnelse for de finkornige gneiser. Dette forslag vakte ufortjent liten oppmerksomhet, og man fortsatte å benytte en rekke forskjellige betegnelser på disse bergarter.

I 1908 tok Högbom leptitbegrepet opp til diskusjon i G. F. i Stockholm gjennom sitt foredrag: »Om en ändring av nomenklaturen för våra granuliter och hälleflintgneiser« (2). Etter foredraget fulgte en lang diskusjon, uten at man kom til noe egentlig resultat. Leptitbetegnelsen som samlenavn for de finkornige gneiser støttet av Sederholm, Törnebohm og J. G. Andersson.

Senere har leptitnavnet vært brukt av mange, med forskjellige definisjoner, mens andre har fortsatt å foretrekke andre betegnelser.

Eskola har i flere avhandinger omtalt leptiter i Finnland, og har definert dem som *finkornige, metamorfe, prekambriske bergarter, som består vesentlig av kvarts og feltspat, med ubetydelige mengder mørke mineraler* (3).

Inntil de aller siste år, er det bare noen ganske få geologer som har foretatt noen større undersøkelse i de prekambriske områder i Norge. — Den første som benyttet betegnelsen leptit her i landet, var W. Werenskiold, som i 1912 utga et geologisk oversiktskart over traktene Setesdal—Ringerike (4). — Arne Bugge har utgitt en rekke avhandlinger fra det Syd-Norske grunnfjell, og det ser etter disse ut til at han såvidt mulig vil unngå leptitnavnet (5).

O. A. Broch (6) har beskrevet en liten del av grunnfjellet på det nordligste av Nesodden, og har her beskrevet leptiter som *finkornige, (metamorfe), fortrinsvis lite skifrige, prekambriske bergarter, med sannsynlig, eller iallfall ikke utelukket, suprakrustal opprinnelse.*

Henrich Neumann (7) har, i motsetning til Arne Bugge, anvendt betegnelsen leptit på Telemarkformasjonens finkornige gneiser. Han har ikke gitt noen definisjon på leptitbegrepet.

Våren 1944 tok Geijer (8) opp leptitbegrepet til diskusjon i G. F. i Stockholm gjennom sitt foredrag »Termen Leptit«. Her gir han følgende definisjon på leptitbegrepet: »*En metamorf (omkristalliserad) suprakrustal bergart, med approximativt granitisk sammansättning, som har en sekundär kornighet emellan 0,03—0,05 mm som undre gräns, och 0,5(—1) mm som övre, bortsett från eventuellt förhandenvarande relikta strökorn.*« Foredraget ble fulgt av en lang diskusjon (8), og jeg skal i det etterfølgende komme tilbake til så vel dette foredraget som flere av innleggene i diskusjonen.

Jeg har selv i de siste fire år arbeidet med geologiske undersøkelser i de prekambriske områder omkring Oslofjord (9) og på Sunnmør. Begge steder viser det seg at leptiter og leptitlignende bergarter er meget utbredt, og dels forekommer over større, sammenhengende omåder. Mangelen på brukbare og definerte begartsnavn har gjort at jeg for en stor del har funnet det nødvendig å lage nye navn.

Prekambriske bergarter med tilnærmet granittisk sammensetning har jeg inndelt etter kornstørrelsen i to hovedgrupper, gneiser og leptiter. Overensstemmende med dette har jeg tillatt meg å samle alle finkornige, metamorfe bergarter med en *tilnærmet* granitisk sammensetning under gruppenavnet leptiter. Med finkornige mener jeg da bergarter med en maksimal midlere kornstørrelse på omkring 0,5 mm, bortsett fra eventuelle relikte strøkorn.

Etter min mening bør man *ved definisjon av leptitene utelukkende ta hensyn til deres kornstørrelse, deres mineralsammensetning, og at de skal være metamorfe.*

I et innlegg etter Geijers foredrag (8), er Backlund kommet til om lag samme resultat, bortsett fra at han ikke vil ha med kravet om at de skal være metamorfe. Han setter opp i 6 punkter de krav som tilsammen av forskjellige geologer er blitt forlangt til definisjon av leptiter:

1. Bestemt lav kornstørrelse.
2. Bestemt mineralsammensetning (tilnærmet granittisk).
3. Bestemt sterk metamorfosegrad.
4. Bestemt opprinnelse (sedimentær eller magmatisk).
5. Nær sammenheng med malmdrag.
6. Suprakrustalbergarter.

Han finner at punkt 1 og 2 uten videre er anvendelige for en definisjon. 3 og 4 mener han at alltid vil bli gjenstand for vidløftige diskusjoner og ulike tolkninger. Om 4 sier han dessuten at det er tvilsomt om det vil være praktisk å knytte en bestemmelse som skulle gi bestemt svar på dette til leptitbegrepet. 5 og 6 vil man først kunne gi svar på etter meget nøyaktige undersøkelser (og ikke alltid da heller). Etter å ha nevnt hvordan de geologer i Sverige og andre land som har befattet seg med leptiter definerer disse, konstaterer han dessuten at 5 i det hele tatt *ikke* later til å være karakteristisk for leptiter.

Som tidligere nevnt, foreslår Geijer (8) at man også skal ha en undre grense for leptitenes kornstørrelse, nemlig 0,03 mm, og han støttes her av flere svenske geologer. De ultrafinkornige bergartene skal etter dette ikke kunne regnes til leptitene, men skilles ut som en egen gruppe, hälleflinter. Dette navnet er et utpreget svensk navn, som ikke kommer til å kunne anvendes utenfor Skandinavia. Dessuten synes jeg navnet virker misvisende. Jeg har derfor hittil ikke villet bruke betgnelsen hälleflint. I diskusjonen etter Geijers foredrag (8) foreslo Quensel med en lignende begrunnelse at man for disse bergarter burde finne et bedre navn, f. eks leptynit. Jeg støtter gjerne dette forslag, men vil dessuten foreslå at man regner leptynitene som en undergruppe innen leptitene. Mange steder er det umulig å avgjøre om en bergart er en leptynit eller en vanlig leptit, på grunn av sterkt varierede kornstørrelse, og svært ofte vil det virke meningsløst å sette en skarp grense mellom en leptynit og en vanlig leptit.

Bergarter med leptitisk mineralsammensetning og med en kornstørrelse på omkring 1 mm eller mindre (ofte svært variert), har jeg betegnet leptitgneiser, når de finnes sammen eller grensende inn til vanlige leptiter. Noe lignende er også gjort av mange andre geologer.

Av det foregående vil det framgå at jeg vil finne det praktisk om vi kunne *inndele prekambriske, metamorfe bergarter med en tilnærmet granittisk sammensetning, og videre leptitene inndeles etter kornstørrelsen i leptitgneiser, vanlige leptiter og leptyniter* (eller et annet navn vi kan bli enige om). *Den maksimale kornstørrelse for leptitenes grunnmasse settes til ca. 1 mm.*

En typisk kvartsit skal altså ikke kunne kalles leptit. Dette ser det for øvrig ut til at alle er enige om. Men Geijer (8) mener, at en bergart som sannsynligvis er en omvandlet sandstein, skal benevnes kvartsit, selv om den fører atskillig feltspat (uansett feltspatinnhold).

Kornstørrelse-skjema for gneis-bergarter.

Alminnelige gneiser	Grov gneis Fin gneis	
		ca. 1,0 mm
Leptiter	Leptitgneiser	
		» 0,5 »
	Vanlige leptiter	
		» 0,03 »
	Leptyniter	

Dette tror jeg vil bli vanskelig å gjennomføre. Det er jo svært alminnelig at man ikke kan uttale noe sikkert om en prekambrisk bergarts opprinnelse.

De mest typiske leptiter er glimmerfattige. Dette er vel også en av de ting alle er enige om. Men det ser ut til å være mange geologer som mener at mere glimmerrike, finkornige bergarter i det hele tatt ikke bør komme inn under leptitbegrepet, tross de jo også svært ofte må sies å ha en tilnærmet granittisk sammensetning.

I sitt foredrag (8) kommer Geijer også inn på disse finkornige, glimmerrike bergarter. Han mener at man må finne en praktisk gruppebetegnelse for disse, for å kunne begrense leptitbegrepet.

Jeg vet ikke hvorfor det egentlig skulle være noe særlig i veien for heller å innbefatte disse bergarter i leptitgruppen. De er meget alminnelige i prekambriske områder, og finnes svært ofte innen utpregede leptitområder. På Nesoddlandet har jeg foreløpig flere steder funnet det praktisk å bruke navnet leptitglimmerskifer, eller glimmerskiferleptit,[1] og under mere detaljert kartlegging — f. eks. i Røyken — har jeg ofte funnet det nødvendig å lage andre navn, sammensatt av leptit og forskjellige mineralnavn. Slike navn vil selvsagt ofte bli nokså tungvinte, men de har den store fordel at de uten videre forteller hva bergarten består av, slik at man slipper en forklaring ved siden av navnet. Et navn som f. eks »epidot- og muskovitførende albitporfyrleptit« blir det jo heller ikke aktuelt å bruke annet enn på særlig detaljerte spesialkart.

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[1] Denne betegnelse benyttet jeg også i mitt foredrag i Norsk Geol. For. 1. febr. 1945. I diskusjonen etter foredraget bemerket S. Foslie at han ikke kunne gå med på betegnelsen »leptitglimmerskifer«, og etter nærmere overveielse er jeg selv også kommet til at det såvidt mulig bør benyttes andre navn, som f. eks. muskovitleptit, biotitleptit o. l.

På et foreløpig oversiktskart over en del av Oslofjordens prekambriske områder (9), har jeg dels funnet det praktisk å anvende stedsnavn sammensatt med leptitnavnet, som Røykenleptit, Tofteleptit og Sproleptit. Men jeg håper dette bare blir midlertidig. Når det blir fastsatt brukbare navn og definisjoner på bergartene innen leptitgruppen vil jeg selvsagt innføre disse istedenfor stedsnavnbetegnelsen.

Det er fremdeles en del geologer som i det hele tatt ikke vil bruke leptitnavnet. Av norske grunnfjellsgeologer ser det ut til å være særlig Arne Bugge som heller vil bruke andre betegnelser på disse bergarter. Leptitene i Telemarkformasjonen betegner han dels »finkornige gneiser«, dels kvartsporfyrer (5). Av det lille jeg har fått høve til å studere disse bergartene, kan jeg ikke se at det skulle være noen særlig grunn til å holde dem utenfor leptitgruppen.

Å betegne leptitene som finkornige gneiser, vil etter min mening medføre mange vanskeligheter. Jeg har til dels brukt denne betegnelsen på bergarter som er for grove til å kalles leptiter. Det kan jo være ganske stor forskjell i kornstørrelse på gneisene, selv om bergarter med en kornstørrelse under 1 mm ikke regnes til denne bergartsgruppe.

Av det som tidligere er skrevet og uttalt om leptitbegrepet, framgår det tydelig at mange vil begrense det sterkt. Såfremt dette blir gjort, tror jeg vi i framtiden kommer til å få mange nye vanskeligheter med nomenklaturen, idet vi da blir nødt til å finne på nye navn for de grupper jeg her har anbefalt at vi henregner til leptitene. Jeg tror den beste løsning vil være at man for disse grupper bruker navn sammensatt med leptit, f. eks. kvartsitleptit, glimmerleptit o. l. Kjenner man med sikkerhet opprinnelsen, settes navnet sammen av den opprinnelige bergart og leptit (også foreslått av Quensel i innlegg etter Geijers foredrag (8). Kan man ikke finne ut noe sikkert om utgangsmaterialets karakter, lager man navnene så godt det lar seg gjøre ut fra det man finner ut om leptitenes mineralsammensetning og struktur.

Mineralogisk Institutt, Oslo, 15. mai 1945.

Litteratur.

1. D. Hummel: »Om Sveriges lagrade urberg«. S. G. U. 1875.
2. A. G. Högbom: »Om en ändring av nomenklaturen för våra granuliter och hälleflintgneiser«. G. F. F. 30, 1908.
3. Pentti Eskola: »On the Petrology of the Orijärvi Region«. Bul. Com. Geol. Finlande, no. 40.
4. W. Werenskiold: »Tekst med geologisk oversiktskart Sætersdalen—Ringerike«. N. G. U. nr. 66.
5. Arne Bugge: »Gradavdelingskartene Flesberg og Eiker«. N. G. U. nr. 143.
6. O. A. Broch: »Ein suprakrustaler Gneiskomplex auf der Halbinsel Nesodden bei Oslo«. N. G. T. bd. 9.
7. Henrich Neumann: »Leptitten i Telemark-Formasjonen«. N. G. T. bd. 22.
8. Per Geijer: »Termen Leptit«. G. F. F. 66, 1944.
9. Chr. Gleditsch: »A Rapid Survey of the pre-Cambrian Areas around the Oslo-Fiord (The Waters Oslo—Son)«. N. G. T. bd. 25.

Trykt 17. januar 1946.

NOTISER

GARDNOSBREKSJEN

Bemerkning til K. Münster Strøms diskusjonsinnlegg, se dette bind p. 262.

Etter mitt foredrag om Gardnosbreksjen ble Münster Strøms diskusjonsinnlegg ufullstendig besvart av grunner som følger nedenfor. Først kan imidlertid kort pekes på følgende:

Ingen iakttagelser i marken tillater en å slutte at breksjeforekomsten er subsirkulært avgrenset, området er altfor overdekket til det. I vest kan en dog se hvordan grensen må gå, og her er dens krumningsradius altfor liten til å passe på en „subsirkel" av dimensjoner som det blir tale om her. En nærmere betraktning viser også at den kegleform som en måtte anta ut fra M. S.'s forutsetning ville være temmelig flat, hvorved — under hensyntagen til terrengets form — de blottninger en ser ville passe ennå dårligere med et subsirkulært tverrsnitt. I øst er den nordlige del av grensen „subrektilineær" (for å bli i stilen). Grensen følger ikke her noen „loddrett" dalside, selv om denne er meget bratt — ca. 45°. (En kan jo for øvrig i alminnelighet slett ikke dømme av kartet om en dalside er (tilnærmet) loddrett.) Sønnenfor (øst for Gardnosgrenda) er riktignok et virkelig stup (70—80°) på vel 100 meters høyde — fra ca. 324 m. o. h. til ca. 460 m. o. h. Dette gjelder altså for en del (halvdel?) av det „subrektilineære" stykke av grensen. Dessverre er jeg først nylig blitt oppmerksom på en vill-ledende tegnefeil i det geologiske kart (se „Gardnosbreksjen i Hallingdal" fig. 2, dette tidsskrift *25*, p. 18). På arbeidskartet ble her foreløpig tegnet en stripe med „breksjeret kvartsit og amfibolit" som så er kommet over på kartet som „kvartsit og amfibolit" mens det skulle vært tegnet som breksje. Jeg var ikke oppmerksom på dette under diskusjonen som derfor virket forvirret. Det er altså ikke så at breksjens grense mot sidebergarten „eksakt" følger en „loddrett" dalside. I foten av stupet er det breksjens grense mot dalsidemorenen som går slik, og det er jo ikke noe påfallende.

En sak for seg er imidlertid spørsmålet hvorvidt den nåværende dagoverflate betegner et snitt gjennom tilførselskanalen eller gjennom krateret. Om en ikke kan godta M. S.'s forutsetninger så er ikke dermed sagt at tanken om at snittet går gjennom krateret er gal. Jeg vil derfor benytte anledningen til å rette den feil jeg har begått ved a priori å utelukke denne mulighet. Spørsmålet blir da hvor dypt terrenget ligger

under overflaten slik den var da eksplosjonen fant sted. Etter det som ble sagt i foredraget (se „Gardnosbreksjen etc." dette tidsskrift *25*, p. 16 osv.) er det mulig at det ikke var særlig dypt. I så fall er det et snitt i krateret som er blottet, og fig. 5 (l. c.) måtte eventuelt bli å rette slik at breksjens grenser konvergerer mot dypet.

Gardnosbreksjen i Hallingdal, a correction. In my paper „Gardnosbreksjen i Hallingdal" (a breccia at Gardnos in Hallingdal) (this journal *25*, pp. 16 et seqq.) I have fallen into an error in leaving out of the discussion the possibility that the outcrop of the breccia may represent a section through a crater. A remark from Kaare Münster Strøm has given me the opportunity of correcting this slip. Thus (cp. fig. 5 l. c.) the walls of the breccia may perhaps converge downwards. — On the same occasion I direct attention to another error, made during the drawing of fig. 2 (l. c.): The strip of „kvartsit og amfibolit" („quartzite and amphibolite") just east of Gardnos does not exist. The outcrop of breccia here abuts on glacial debris (valley-side moraine).

Olaf Anton Broch.

NORSK GEOLOGISK FORENING

MØTE 285. TORSDAG 30. NOVEMBER 1944

Til stede 26 medlemmer, 7 gjester.

INNVALG

266. Assistent HARALD MAJOR, Innheradsvegen 129, Trondheim.
Etter forslag av Th. Vogt og K. Münster Strøm.

Formannen meddelte at fra og med bind 24 vil Norsk geologisk tidsskrift bli sendt ut til medlemmene i beskåret tilstand. (En del medlemmer som har uttalt ønske om det vil fortsatt få heftene uoppskåret.)

FOREDRAG

TOM. F. W. BARTH: *Christensenit, en ny mineralserie.*

Trykt i nr. 20 av serien Scientific Results of the Norwegian Antarctic Expeditions 1927—1928 et sqq., Instituted and Financed by Consul Lars Christensen, utgitt av Vid.-Akad. Oslo 1945.

K. Kristoffersen uttalte seg etter foredraget om muligheten av å framstille christensenit kunstig.

J. A. DONS: *Resultater av den geologiske kartlegging i Ullernåsen.*

Ullernåsen ligger få kilometer vest for Oslo som en skogkledd kolle omgitt av villabebyggelse. Helt siden 1820-årene da Keilhau oppdaget området som finnested for „essexit“ er åsen og dens bergarter ofte blitt beskrevet og undersøkt av geologer. Særlig må nevnes prof. W. C. Brøgger og amanuensis P. Schei.

Det som har vanskeliggjort tidligere arbeider er dårlige kart og meget overdekking.

Studiet av Ullernåsen samt et tilstøtende breccie-område er nå gjenopptatt av foredragsholderen. Kartgrunnlaget er de helt utmerkede nye kart fra Akers Oppmålingsvesen (1 : 1000 og 1 : 5000). Forskjellige sprengningsarbeider har gitt enkelte verdifulle snitt.

Etter å ha beskrevet de topografiske forhold og omtalt tidligere geologiske arbeider om Ullernåsen forklarte foredragsholderen ved hjelp av et stort kart, to profiler og et blokkdiagram de foreløpige resultater av den geologiske kartlegging.

For å kunne påvise eventuelle forkastninger, er sedimentene detaljkartlagt. Kontaktmetamorfose fra åsens dyperuptiver og ganger har vanskeliggjort arbeidet. Det finnes 3 synklinaler og 3 antiklinaler, med stadig vekslende lag fra 4 aα til 4 bγ. Dypbergarten er trengt opp i en synklinal. Lagstillingen er de aller fleste steder steil.

Kart og profiler viser med all tydelighet at den dype Mærradalen som i ca. S—N retning skjærer seg gjennom Ullernåsen følger en forkastningslinje. Den vestre side er forskjøvet ca. 200 m mot nord og hevet ca. 50 m i forhold til den østre side. Kvarts- og kalkspat-breccier, strukturer innen den forkastede „essexitmasse", små parallell-forkastninger m. m. viser det samme.

Følges forkastningens retning mot syd fortsetter den i den store forkastning langs Oslofjordens østside. Ved Hvaler er det påvist at denne har en tilsvarende forskyvning av den vestre side i forhold til den østre. H. Cloos mener teoretisk og etter feltgeologiske iaktagelser å kunne vise at ved de fleste forkastninger i Oslo-området er vestre side skjøvet sydover i forhold til østre. Dette står altså i strid med de her nevnte forkastninger, samt en stor forkastning i. Sandviksdalen.

En stor glimmersyenitporfyrgang som antas å ha sammenheng med forkastningen i dypet, kan følges fra Nesodden over forskjellige øyer, opp gjennom Ullernåsen til et akeritmassiv i Holmenkollåsen. Dette, sett i sammenheng med S—N strukturen i det underliggende grunnfjell, sansynliggjør sammenhengen mellom Mærradalsforkastningen og den store permiske forkastning i Oslofjorden. Glimmersyenitporfyrgangen forgrener seg i Ullernåsen. En analog forgrening av forkastningen finnes på tilsvarende punkt i strøkretningen. Gangen har mørke, basiske ganggrenser, sansynligvis oppstått ved differensiasjon i selve gangspalten som følge av avkjøling langs selve gangveggen.

Dypbergarten som tidligere var antatt å være essexit viser seg å ha syenitdioritisk sammensetning tilsvarende kauaiit. Videre forekommer akeritiske typer og husebyit av nefelinsyenitisk sammensetning. Differensiasjonen synes å ha foregått i forskjellige retninger innen de 3 massiver i Ullernåsen.

Innen området er funnet 4 forekomster av eruptivbreccier tilsvarende Werenskiolds Jar-Øraker breccier. Videre en 3 cm. bred brecciegang med bl. a. grunnfjellstykker på 2 ½ cm. diameter.

Analysering og mikroskopering av bergartene samt kartlegging av nå avsperrede områder står ennå igjen.

I ordskiftet etter foredraget deltok Holtedahl, Størmer, Werenskiold, Chr. Oftedahl, A. Bugge og foredragsholderen.

EGIL SÆTHER: *Undersøkelser over eruptivområdet nord for Oslo.* Trykt i bind 25, p. 427.

I ordskiftet etter foredraget deltok Holtedahl, Barth, Chr. Oftedahl, J. A. Dons, C. Bugge, A. Bugge og foredragsholderen.

GENERALFORSAMLING OG MØTE 286.
TORSDAG 1. FEBRUAR 1945.

Til stede 27 medlemmer.

GENERALFORSAMLING

ÅRSMELDING FOR 1944

Siden forrige generalforsamling er 4 medlemmer avgått ved døden: Carl Wiman, Sverre Blekum, A. Nummedal, R. Hvoslef.

I samme tidsrom er innvalgt 2 nye medlemmer:

265. Ingeniør Aslak Kvalheim, Geologisk museum, Oslo, $^{23}/_{3}$ 1944.
266. Assistent Harald Major, Innheradsvn. 129, Trondheim, $^{30}/_{11}$ 1944.

Medlemstallet er nå 151, derav 73 livsvarige og 78 årsbetalende.

Det er holdt 6 møter med et samlet frammøte av 163 personer, samt 1 ekskursjon med 17 deltagere.

Av tidsskriftet er utkommet bind 22, h. 3—4 og hele bind 23.

UTDRAG AV REGNSKAP FOR 1944

Inntekt.

Beholdning overført fra 1943:		
I kasse	kr. 259,18	
I bank	„ 29987,76	kr. 30246,94
Innkommet medlemskontingent:		
For 1941 fra 1 medlem	kr. 10,00	
„ 1942 „ 11 medlemmer	„ 110,00	
„ 1943 „ 17 „	„ 170,00	
„ 1944 „ 47 „	„ 470,00	kr. 760,00
Renter fra Livsvarige medlemmers fond		„ 171,08
Renter fra Berg- og steinindustriens fond		„ 54,37
Tilskudd fra staten		„ 400,00
Tilskudd fra Sulitjelmafondet		„ 2600,00
Abonnement og salg av tidsskriftet		„ 5202,00
Annonser i tidsskriftet		„ 1875,00
Renter av bankinnskudd		„ 30,10
		kr. 41339,49

Utgift.

Tidsskriftet:		
Trykning bind 22, hefte 1—2	kr. 2012,00	
— „ „ „ 3—4	„ 2882,00	
— „ 23, „ 1—3	„ 3698,00	
		kr. 8592,00
Trykning av kart til bind 25		„ 618,75
Klisjeer til bind 23, hefte 1—3	kr. 451,01	
— „ „ 23, „ 4	„ 226,13	
— „ „ 24, „ 1—2	„ 271,46	
		kr. 948,60
Arbeidshjelp, porto og skrivesaker		„ 564,55
Møter og representasjon		„ 105,00
Rabatt ved salg av tidsskriftet		„ 1338,02
Overført til Berg- og steinindustriens fond. *Urørlig kapital*		„ 20000,00
Beholdning overført til 1945:		
I kasse	kr. 96,78	
I bank	„ 9075,79	
		kr. 9172,57
		kr. 41339,49

Livsvarige medlemmers fond.

		Inntekt	Utgift
1944 $^{1}/_{1}$ Fondets kapital:			
Obligasjoner	kr. 3500,00		
Bankinnskudd	„ 2140,00		
Innbet. livsv. kontingent	„ 300,00		
Urørlig kapital	kr. 5940,00		
Renter av obligasjoner i 1944		kr. 126,00	
Renter av bankinnskudd i 1944		„ 50,08	
Forvaltning av obligasjoner			kr. 5,00
Overført renter til ordin. budgett			„ 171,08
		kr. 176,08	kr. 176,08

Berg- og steinindustriens fond.

Fondets kapital	kr. 20000,00
Renter for 1944	„ 43,27
Overført renter til ordin. budgett	„ 43,27

Urørlig kapital kr. 20000,00

Regnskapet er revidert, Oslo, $^{15}/_{1}$-1945. S. Foslie, J. Helverschou.

REUSCHMEDALJEN

Cand. real. Per Holmsen ble tildelt Reuschmedaljen for sin avhandling: „Geologiske og petrografiske undersøkelser i området Tynset — Femunden.“ N. G. U. skr. Nr. 158, 1943.

VALG

Styre for 1945.

Formann: .... Münster Strøm
Sekretær: .... Chr. Oftedahl
Redaktør: .... I. Oftedal
Styremedlemmer: .... Heintz
Strand
Varamenn:...................... Jens Bugge
Rosenlund

Revisorer for 1945: Helverschou og Foslie med H. H. Smith som varamannn.

Reuschmedaljekomité for 1945: Foslie og Holtedahl.

LOVFORANDRING

§ 8. „...gode rentebærende papirer“ forandres til „...gode verdier“.
„...fondets renter“ forandres til „...fondets avkastning“.

Foslie bemerket at tilsvarende forandring bør foretas i statuttene for Berg- og steinindustriens gavefond. Helverschou bemerket at med „gode verdier“ må forståes inntektgivende verdier.

MØTE 286.

Siden forrige møte var H. H. Horneman avgått ved døden. Minnetale vil bli holdt av C. Bugge på neste møte.

INNVALG

267. Bergingeniør Fritz Øien, Killingdal Grube, Reitan, — etter forslag av Jens Bugge og Münster Strøm.
268. Disponent Erik Sverre jr., Drammensvn. 163, Oslo, — etter forslag av Barth og Chr. Oftedahl.

FOREDRAG

Chr. Gleditsch: *Om leptit-begrepet.* Trykt i dette bind p. 244.

I diskusjonen etter foredraget deltok A. Bugge, Jens Bugge, C. Bugge, Werenskiold, Foslie, Barth, Sæther og foredragsholderen.

Statsgeolog ARNE BUGGE uttalte vedrørende anvendelsen av leptitnavnet at han i 1928 hadde brukt det om Telemarkformasjonens finkornige gneiser således som det tidligere var benyttet av Werenskiold. Under en senere kartlegging av rektangelbladet Kongsberg viste det seg at det ikke var mulig å atskille omvandlede kvartsporfyrer fra mylonittiserte granitter og feltspatkvartsitter. Leptitnavnet kunne altså ikke anvendes utelukkende som betegnelse på sikre dekkebergarter. Han sluttet da med å benytte lepetitbetegnelsen for de finkornige bergarter i Telemarkformasjonen særlig da navnet i Sverige mer og mer ble innarbeidet som bergartsnavn i den malmførende leptitformasjon.

Selv om man nok kan si at leptitnavnet er petrografisk definert, således at man kan anvende det utenfor den egentlige leptitformasjon, så vil vel de fleste geologer og bergingeniører, når de hører om leptit, straks anta at det er tale om en bergart som finnes i den malmførende svenske leptitformasjonen. Navnet er svensk i opprinnelse og anvendelse og det er i Sverige hevdet gjennom så fremragende beskrivelser fra den egentlige leptitformasjon, at det vil være riktig at man også i framtiden knytter det til denne formasjon og ikke utvider det til å betegne finkornige grunnfjellsbergarter i sin alminnelighet.

Skulle det være et ønske fra de svenske geologers side, at navnet blir anvendt også i andre land og om finkornige bergarter i hvilkensomhelst grunnfjellsformasjon uten hensyn til om de er malmførende, hadde Bugge selvfølgelig intet å innvende mot dette, om han enn ville beklage at navnet mistet sin spesielle klang.

S. FOSLIE: Når vi her drøfter anvendelsen av betegnelsen leptit på norske bergarter må vi ha for øye, at det i sin opprinnelse er et svensk navn, at svenskene har prioriteten og dermed retten til å avgjøre hva dette navn skal omfatte.

Hvis derfor svenskene selv var enige i spørsmålet var det ingen grunn til å diskutere det her hos oss. Dette synes imidlertid ikke å være tilfellet og det var vel nettopp grunnen til at overdirektør Geijer fant seg foranlediget til å holde et foredrag og diskusjonsmøte om dette spørsmål. De forskjellige medarbeidere ved S. G. U. viste en tendens til å legge hver sin private betydning i begrepet leptit.

Så lenge man holder seg til Bergslagen synes saken å være klar. Leptitbetegnelsen er her nøye knyttet til bergarter i en malmførende formasjon. Når man imidlertid kommer utenfor dette område, særlig opp mot Nord-Sverige og til ikke malmførende områder er det at uenigheten begynner.

En ting må man imidlertid være klar over. Når det på enkelte hold i Sverige forlanges, at betegnelsen leptit bare skal anvendes innenfor visse malmførende områder, så måtte i tilfelle leptit utgå av den petrografiske nomenklatur og gå over til en formasjonsbetegnelse. Man kan ikke knytte en slik betingelse til et petrografisk navn. Av diskusjonen i Geologiska Föreningen synes på den annen side å framgå, at det overveiende antall svenske geologer ønsker leptit opprettholdt som et petrografisk begrep. Dermed skulle altså forutsetningen være til stede til å anvende betegnelsen på passende bergarter også utenfor Sverige.

Vi har her i Norge en nokså utbredt gruppe av finkornige, sterkt eller helt omkrystalliserte kvarts-feltspatbergarter med forholdsvis lite mørke mineraler og av usikker opprinnelse.

I de tilfelle hvor bergartens opprinnelse med sikkerhet kan konstateres, — som sediment, som kvartsporfyr eller som facies av dyperuptiv — kan man alltid finne en passende betegnelse for den, men så lenge opprinnelsen er usikker er det, at vi føler et behov for å bruke samlebetegnelsen leptit, fordi ingen annen tilnærmelsesvis treffer blinken.

Å føye prefikser til navnet, som f. eks. glimmer-leptit, skulle det ikke være noe iveien for, men når Gleditsch i sitt inndelingsforslag også bruker ordet „glimmerskifer-leptit“ kommer man ut på viddene. Glimmerskifer skal etter sin definisjon være en i hovedsaken feltspatfri bergart, mens leptit nettopp skal være feltspatførende, så å koble dem sammen ville bare føre til uklarhet.

O. HOLTEDAHL: *Bemerkninger om forholdet mellom Telemark- og Kongsberg-Bamble-formasjonene.*

Vedr. professor Holtedahls foredrag framkom ARNE BUGGE med en lengere uttalelse. Der ble dog ikke anledning til å behandle de av Holtedahl reiste spørsmål i hele sin bredde. Vedr. Holtedahls antagelse at rivningsbreksjen mellom Kongsberg-Bamble formasjonen og Telemarkformasjonen muligens har fått sitt særpreg i kaledonisk tid, uttalte Bugge: Den store rivningsbreksje som har en bredde opp til 300 m er dannet i 3 forskjellige epoker som er karakterisert ved: 1. mylonittisering, 2. kvartsganggjennomveving, 3. langsgående, ofte leirfylte spalter.

1. Mylonittisering finnes i den store rivningsbreksjen i opp til 300 m bredde, og svinner jevnt hen innover i Kongsbergformasjonen. Parallellbreksjer i Kongsbergformasjonen er også mylonittiseret. Mylonittisering forekommer kun i Kongsberg-Bambleformasjonens bergarter. Telemarkformasjonens „granit“ samt Birkeland og Flågraniten er kun oppbrutt, men ikke mylonittisert.

2. Kvartsganggjennomveving forekommer praktisk talt i alle grunnfjellsbreksjer.

3. Langsgående leirfylte spalter sees i alle grunnfjellsbreksjer. Dessuten skjærer disse spalter fra grunnfjellsbreksjene som forkastningsspalter inn i Oslofeltets bergarter der hvor rivningsbreksjen har retning mot dette. De hevninger og senkninger som i post-permiske tid har foregått langs spaltene har naturligvis intet å gjøre med de forskyvninger som pågikk i Kongsberg-Bamble og Telemarkformasjonen i svionisk og gotisk tid.

Man har altså i rivningsbreksjen aldersfølgen:

1. De mylonittiserede breksjer er eldst, idet de kun forekommer i Kongsberg-Bambleformasjonens bergarter.

2. Kvartsgjennomvevingen har inntruffet etter Grimstad-Flå-granitens størkning, men da kvartsbreksjene avskjæres av Oslofeltets skifre og eruptiver, må de være av prekambrisk alder.

3. De langsgående spalter skjærer inn i Oslofeltets skifre og eruptiver og må altså være av post-permisk alder.

Holtedahl regnet i sitt foredrag også med muligheten av at bergartene i Bamble og Tørdal var foldet i kaledonisk tid og at de hørte til de samme foldete lag. Bugge kunne ikke finne noe grunnlag for en sådan antagelse. Samme strøkretning og overensstemmelser i bergart-typer er ikke tilstrekkelig grunnlag for å sammentegne fjerntliggende bergarter som i samme periode foldede lag. Der er meget som må overveies når man parallelliserer. Det var et særdeles omfattende spørsmål som Holtedahl reiste ved å anta at der er kaledoniske foldninger i Telemark og Bambleformasjonen, og Bugge fant det nødvendig for å klargjøre sin oppfatning å parallellisere med det sydsvenske grunnfjell.

Han anbefalte at man i det norske grunnfjell benytter den av N. Magnusson foreslåtte syklusinndeling. Man får derved et sammenligningsgrunnlag, således at man kan få en samlet oversikt over det norske og svenske grunnfjell. Han kunne ikke se at man kunne nå fram til en riktig eller praktisk inndeling i grunnfjellet, når man ved hjelp av strøkretninger søker å binde det sammen i forskjellige fjellkjeder.

Arendal-Bambletraktene er f. eks. forsøkt medregnet i svekofenniderne og Bambleformasjonen er langs Sørlandskysten tegnet med øst svakt nordlig strøkretning. Dette er ikke riktig. Man finner en avgjort nordøstlig strøkretning. På Modum, hvor bergartene nødvendigvis må regnes til Bambleformasjonen har man nord-sydlig strøk og ikke svekofennidernes øst-vestlige.

Området fra Kristiansand og østover langs breksjen har ikke som enkelte oversiktskarter viser nord svakt vestlig strøkretning, men nord svakt østlig. — Det lar seg derfor ikke gjøre, således som H. Backlund har forsøkt det å søke å innpasse de norske arkeiske formasjoner i de gamle fjellkjeder, når man vil søke å inndele dem i overensstemmelse med strøkretningene. Man kommer til en langt klarere inndeling ved å holde seg til syklusinndelingen og regne Kongsberg-Bambleformasjonen til den svioniske syklus og Telemarkformasjonen til den gotiske.

Fra den vest-svenske gneis er der som bekjent en ubrutt sammenheng med bergartene vestover til Moss og Nesodden og videre nordover til Mjøsa langs den av Magnusson beskrevne mylonittsone.

Bugge nevnte at han hadde fulgt „Mylonittsonen" til den store breksje ved Eina hvor den avskjæres. Da der ikke kan sees noen grunn til å regne den vanlige gneisgraniten på vestsiden av breksjen til noen annen formasjon, føres man her skrittvis over fra den vestsvenske gneis til „Telemarkgraniten" og må regne denne — i likhet med vestsvensk gneis — til den gotiske syklus med den absolutte aldersbestemmelse fra Moss til noe over 800 mill. år.

Den store ansamling av vel bevarte suprakrustalbergarter i Telemarken med de underliggende bunngneiser kan etter et sådant resonnement sammenlignes med Gilbergaskålen. Forøvrig finner man i hele det store „Telemarkgranit“ felt en regional migmatisering og tektonisering som har stor likhet med forholdene i den vestsvenske gneis.

I Kongsberg-Bambleformasjonen har man ved Arendal en absolutt aldersbestemmelse 1047 mill. år, som viser at denne formasjon må henregnes til den svioniske syklus.

I Bambleformasjonen er det et karakteristisk trekk at graniter og gabroide bergarter sonart er omgitt av pegmatitganger, kalkspatganger, dolomitganger, forskjellige mineralganger og hydrotermale og pneumatolytiske mineraliseringer, som må være dannet under og umiddelbart etter granitenes og de gabroide bergarters intrusjonsperiode. Disse bergarter, ganger og mineraliseringer har iflg. foranstående resonnement sin opprinnelse i svionisk tid og viser ikke spor av de regionale migmatiseringer og tektoniseringer som har rammet den nærliggende „Telemarkgranit“ i gotisk tid. Kaledoniske foldninger kan da naturligvis heller ikke tenkes.

Det var videre Bugges hensikt å nevne en del iakttagelser han under sine reiser hadde gjort i det syd-norske grunnfjell for med den gitte oversikt som grunnlag å utrede i hvilken grad han mente grunnfjellet var påvirket av den kaledoniske fjellkjedefoldning. Der ble dog ikke anledning til dette, hvorfor han må søke å oppta dette spørsmål ved en annen anledning.

MØTE 287. TORSDAG 1. MARS 1945.

Til stede 24 medlemmer, 2 gjester.

Formannen meddelte at det gjennom bergingeniør Helverschou var innkommet en gave på 2500 kr. fra Christiania Portland Cementfabrik, samt et første årlig bidrag av 100 kr. fra Skaland Grafitverk.

Foreningen kan feire 40-årsjubileum for sin stiftelsesdag 18. februar 1905 og sitt første møte 26. mars 1905.

C. BUGGE holdt minnetale over bergingeniør H. H. HORNEMAN. Den blir trykt i bind 26.

FOREDRAG

TRYGVE STRAND: *Strukturundersøkelser i Bygdinkonglomeratet.* Trykt i dette bind p. 14.

I diskusjonen etter foredraget deltok Werenskiold, Per Holmsen, Chr. Oftedahl, Jens Bugge og foredragsholderen.

HANS RAMBERG: *Om granitisering og dannelse av pegmatiter ved diffusjon i faste bergarter.*

I diskusjonen etter foredraget deltok Werenskiold, Barth, Kristoffersen, Jens Bugge, Sæther, Foslie, Dietrichson og foredragsholderen.

I anledning av 40-årsjubileet var aftensbordet festligere enn vanlig, og det ble holdt taler av Münster Strøm, C. Bugge, Holtedahl, Barth, Werenskiold og Størmer.

EKSTRAORDINÆR GENERALFORSAMLING OG MØTE 288.

ONSDAG 11. APRIL 1945.

Til stede 17 medlemmer.

GENERALFORSAMLING

FORANDRINGER I LOV OG STATUTTER

Lovens § 7. „. . . styre og 2 revisorer med varamenn“ forandres til „. . . styre med 2 varamenn samt 2 revisorer med varamenn“. Det tilføyes: „Hvert personlig medlem har én stemme.“

Lovens § 9. Det tilføyes: „Det er adgang for alle, også for bedrifter, å tegne seg som støttende medlemmer av foreningen ved årlige bidrag eller bidrag en gang for alle.“

Statutter for Berg- og steinindustriens gavefond, § 2. „Dets midler skal anbringes således som for offentlige stiftelsers og legaters midler bestemt“ forandres til: „Komitéen anbringer fondets midler.“ „renteavkastning“ forandres til „avkastning“. Etter „regnskap“ tilføyes: „revidert av foreningens revisorer“.

MØTE 288.

INNVALG

269. Konsulent S. K. HØEGH-OMDAL, Utgarvn. 12, Jar, — etter forslag av Helverschou og Kvalheim.
270. Cand. mag. LEIV GJESSING, Geologisk museum, Oslo, — etter forslag av Münster Strøm og Chr. Oftedahl.

FOREDRAG

STEINAR FOSLIE: *Om hastingsiter og hornblende fra epidotamfibolit-facies.* Trykt i bind 25, p. 74.

I ordskiftet etter foredraget deltok Strand, Chr. Oftedahl og foredragsholderen.

PER HOLMSEN: *Et leirfall ved Lade, Trondheim.*

Foredraget framkalte bemerkninger av Helverschou, G. Holmsen og foredragsholderen.

MØTE 289. ONSDAG 2. MAI 1945.

Til stede 19 medlemmer, 1 gjest.

INNVALG

271. Ingeniør JOHS. GRENNESS, Geologisk museum, Oslo, — etter forslag av Oftedal og Kristoffersen.

272. Dr. CARL M:SON MANNERFELT, Generalstabens Litografiska Anstalt, Vasagt. 16, Stockholm, — etter forslag av Münster Strøm og Tore Sund.

Det ble bekjentgjort at foreningen ville foreta en ekskursjon til „Bærumskalderaen".

Formannen meddelte at det er blitt dannet en geologisk klubb i Trondheim og at en liknende klubb vil bli dannet i Bergen når Kolderup kommer tilbake.

FOREDRAG

TOM F. W. BARTH og CHR. OFTEDAHL: Korte meddelelser om *to usedvanlige lavatyper i den laveste del av de permiske effusiver ved Horten.* Blir trykt i Vid-Akad. Skr. (Barth) og N. G. T. bind 26 (Oftedahl).

I diskusjonen etterpå deltok Sæther, Broch, Münster Strøm, Holtedahl, og Barth.

O. A. BROCH: *Gardnosbreksjen i Hallingdal.* Trykt i bind 25. p. 16.

I diskusjonen etter foredraget deltok Barth, Chr. Oftedahl, Størmer, Münster Strøm og foredragsholderen. Se også side 250.

MÜNSTER STRØM gjorde oppmerksom på at kartbildet tyder på en traktformet, ikke en loddrett begrensning av breksjen. Hallingdalens bratte vestvegg skjærer rettlinjet over det ellers subsirkulære kartomriss. Det vil være mot sannsynligheten at vestveggen eksakt skal falle sammen med et loddrett begrensningsplan som fortsetter i dypet. Kartbildet forklares derimot helt om vi tenker dalens vestvegg bare som en erosjonsbegrensning. Breksjen blir da en kjegle med basis i fjellplatået overskåret av dalens vestvegg som et nesten loddrett plan. Dette plan skjærer den subsirkulære basis som en korde. (Alle uttrykk å forstå rent geometrisk.)

IVAR OFTEDAL: *Om planeten Jordens spesifikke egenskaper.*

Det er alminnelig antatt at de større planetene i sin tid er avspaltet av Solen som følge av en bestemt prosess eller begivenhet; særlig den ting at de alle er „rettløpende" og holder seg nær ett og samme plan gjennom Solen kan vi vanskelig tenke oss noen annen rimelig forklaring på. Vi går derfor ut fra at de større planetene og deres store måner kjemisk har samme eller liknende sammensetning som Solen — og for øvrig også flertallet av de andre stjernene. Direkte kan vi ikke få undersøkt planetenes og månenes kjemiske sammensetning, men vi mener å ha fått et noenlunde riktig begrep om den universelle gjennomsnitts-sammensetningen gjennom studiet av jordskorpen, meteorittene og stjernespekterne.[1]

Som grunnlag for en sammenlikning av forhold på Jorden med tilsvarende på andre kloder i planetsystemet er her i tabellform sammenstillet en rekke data som nå regnes for vel underbygget. I Tabell 1 er enheten for tallstørrelsene over alt den verdi som gjelder for Jorden. Klodene er ordnet etter avtakende masse.

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[1] Se f. eks. V. M. Goldschmidt, Geochem. Verteilungsgesetze IX, Vid.-Akad. skr. Oslo, I, 1937, No. 4.

Tabell 1.

	m	r	m/r^3	m/r^2	$\sqrt{m/r}$	$1/A^2$	*Atm.*
[Solen]	[331950.]	[109.1]	[0.256]	[27.89]	[55.17]		
Jupiter	316.94	11.0	0.24	2.64	5.35	0.037	+ + +
Saturn	94.9	9.0	0.13	1.17	3.25	0.011	+ + +
Neptun	17.16	4.1	0.25	1.03	2.04	0.001	+ + +
Uranus	14.66	3.8	0.26	1.00	1.95	0.003	+ + +
Jorden	1.00	1.00	1.00	1.00	1.00	1.000	+ +
Venus	0.81	0.97	0.88	0.85	0.91	1.920	+ +
Mars	0.11	0.53	0.72	0.38	0.45	0.430	+
Merkur	0.04	0.39	0.70	0.27	0.32	6.600	—
Jupitermåne III	0.026	0.40	0.39	0.16	0.25	0.037	—
Saturnmåne VI	0.023	0.34	0.58	0.20	0.26	0.011	—
Jupitermåne IV	0.016	0.41	0.24	0.10	0.20	0.037	—
Jupitermåne I	0.013	0.29	0.51	0.14	0.21	0.037	—
Jordens måne	0.012	0.27	0.60	0.17	0.21	1.000	—
Jupitermåne II	0.008	0.25	0.53	0.11	0.18	0.037	—

m er massen, r den midlere radius.

m/r^3 gir den midlere tetthet. I vanlig mål er enheten 5.52.

m/r^2 gir den midlere verdi av tyngdens akselerasjon på klodens overflate.

$\sqrt{m/r}$ gir den midlere verdi av „løsrivningshastigheten“ (velocity of escape) på klodens overflate; det er den minimalhastighet et projektil må ha for å komme fri av klodens gravitasjonsfelt. I vanlig mål er enheten 11.2 km/sek.

A er klodens middelavstand fra Solen. $1/A^2$ gir oss derfor den relative midlere intensitet av den mottatte solstråling.

Atm. I denne kolonne er angitt om en atmosfære kan konstateres på kloden eller ikke; det er her også forsøkt en gradering.

I den neste tabell er oppført de kloder som roterer raskt og har en godt målbar flattrykking ved polene.

Tabell 2.

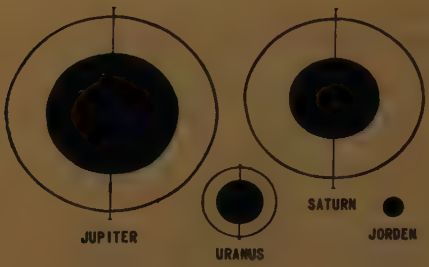
	$p/86164$	f	q	$f.q$	r_1/r
Jorden	1.00	0.00337	290	0.98	1.00
Mars	1.03	0.00521	220	1.15	0.90
Jupiter........	0.41	0.0650	12	0.78	0.62
Saturn	0.43	0.105	7	0.73	0.51
Uranus	0.45	0.055(?)	16	0.88(?)	0.64

p er rotasjonstiden i sekunder, $p/86164$ i stjernedager.

f er meridianellipsens flattrykking ([a—b]/a).

q er forholdet mellom tyngdekraft og sentrifugalkraft ved planetens ekvator ($p^2 d \cdot G/3\pi = 0.708 \cdot 10^{-8} \cdot p^2 d$, hvor G er gravitasjonskonstanten og d planetens midlere sp. v. i vanlige enheter).

I neste kolonne står produktet $f.q$. Dette produktet er avhengig av massefordelingen i klodens indre. For konstant tetthet gjennom hele kloden (inkompressibel veske) har det sin største verdi, omkring 1.25, og for en klode hvor praktisk talt hele massen er samlet i sentrum har det sin minste verdi, omkring 0.5.



I siste kolonne er r_1 den radius planeten ville hatt hvis dens masse var like tett sammenpakket som Jordens. r_1/r er forholdet mellom denne radius og den virkelige. De vesentlige data fra denne kolonne er illustrert på figuren, som også viser planetenes dimensjoner i riktig forhold.

Vi ser av Tabell 1 at Jorden har den største midlere tetthet av alle klodene. Venus kommer ikke langt etter, og for øvrig har Mars, Merkur og et par av månene tettheter som ligger vesentlig over halvdelen av Jordens. De andre klodene har bare brøkdeler av Jordens tetthet. For de store ytre planetene er den lave midlere tetthet etter alt å dømme en følge av enormt tykke skyfulle gasshylstere som hindrer oss i å iaktta planetenes egentlige overflater. Figuren viser hvor tykke disse atmosfærene måtte være hvis planetene (atm. tetthet satt=0) hadde Jordens midlere tetthet. Atmosfærens høyde opp til de øverste skylagene ville på Jupiter være 27000 km og på Saturn 30000 km. Om atmosfærene virkelig er så tykke vet vi jo ikke. Men at disse kjempeplanetenes masser er sterkt konsentrert mot sentrum framgår også av verdiene for produktet $f.q$ (Tabell 2), som på den annen side tyder på at Mars har en mindre utpreget massekonsentrasjon mot sentrum enn Jorden. Vi kan merke oss at for Jupiter og Saturn (og Uranus) fører både de beregnete atmosfæretykkelser og $f.q$-verdiene til det samme resultat: at Saturn har den sterkeste massekonsentrasjon mot sentrum. At Jupiters, Saturns, Uranus' og Neptuns atmosfærer har tykkelser av den nevnte størrelsesorden er vel det sannsynligste, selv om de lavere atmosfærelag på grunn av trykket kanskje har en tetthet som kan modifisere resultatet.

Mens disse kjempeplanetenes lave midlere tetthet altså kan forklares ved å anta at deres atmosfærer når titusener av kilometer ned under den synlige overflaten er en liknende forklaring neppe mulig når det gjelder Jupitermånene III og IV, som også har meget lav midlere tetthet, den siste bare 1.4 ganger vannets. Disse månene er større og tyngre enn Jordens måne, men „løsrivingshastigheten" er for alle tre omtrent den samme (2.8, 2.2, og 2.4 km/sek), og det er vesentlig denne som betinger om kloden er istand til å fastholde en atmosfære eller ikke, da det jo er spørsmål om hvorvidt molekylenes hastigheter i atmosfæren når denne grenseverdi. Lovene for molekylenes hastigheter i gassarter kjenner vi og kan derfor — for en antatt temperatur — regne oss til om en klode kan fastholde en atmosfære og hvilke av de vanlige gassarter denne i tilfelle kan bestå av. Imidlertid har vi en del observasjonsmateriale å holde oss til, se Tabell 1. Mars er den minste kloden som sikkert har atmosfære, og den må være meget tynnere enn jordens. At Merkur, som det ser ut til, helt mangler atmosfære kan delvis komme av den sterke solvarmen (Tabell 1, $1/A^2$), som ville øke gassmolekylenes hastigheter, men alt i alt kan vi tydeligvis si at kloder som har løsrivingshastigheter under ca. 1/3 av Jordens mangler atmosfærer. Jordens måne er sikkert uten atmosfære, og for de andre månene peker observasjonene avgjort i samme retning. Vi må derfor gå ut fra at Jupitermånene, særlig III og IV, består av ikke gassformet materiale med meget lav egenvekt, og i rekken fra Merkur og nedover i Tabell 1 har vi derfor sterke svingninger i den midlere tetthet. Kuhn og Rittmann[1] mener at planetsystemets kloder består vesentlig av samme stoff („solarmaterie"), som er komprimert, sterkere jo større klodens masse er, og dette vil jo si, at tettheten stiger med massen. Etter denne teorien burde Jupitermåne III hatt en midlere tetthet av ca. 0.65 (mot 0.39), og Jupitermåne IV ca. 0.60 (mot 0.24). (Jeffreys har tenkt seg at disse månene for en stor del består av materialer som is og frossen kullsyre, da alminnelige bergarter er altfor tunge til å gi en så lav midlere tetthet.) Liksom vi antar at planetene er oppstått av solen må vi anta at i hvert fall de fleste av månene er oppstått av sine sentralkloder. Dette gjelder særlig for de store månene til Jupiter og Saturn, da de beveger seg praktisk talt i sentralklodens ekvatorplan og samme vei som denne roterer. Da, som vi har sett, de fire store Jupitermånene neppe kan være likt sammensatt må vi altså slutte at stoffet i Jupiter har vært differensiert før månene ble avspaltet, d. v. s. etter alt å dømme på et meget tidlig stadium i planetsystemets historie. Er differensiasjon mulig i en planetmasse som er i stand til å avspalte måner så burde den også være tenkelig for selve solmassen. Den moderne teori for stjernenes konstitusjon må riktignok regne med udifferensiert stoff i Solens indre, men det kan godt være at dettes sammensetning avviker fra solatmosfærens slik som vi kjenner den fra studiet av spektrene; Eddington har spesielt betonet dette. En må derfor anse det som en mulighet at planetene fra begynnelsen av har fått en noe forskjellig middelsammensetning. For kloder som ikke eller bare delvis er istand til å fastholde en atmosfære må vi regne med en forandring i sammensetningen etter at de ble til, da jo flyktige stoffer som kommer fram til overflaten etter hvert må forsvinne ut i rommet. Men noen vesentlig rolle kan dette neppe ha spilt, da dannelsen av et fast overflateskikt i det vesentlige må ha stoppet denne prosessen. Dessuten burde den vel snarere føre til en *øking* av klodens midlere tetthet, i strid med det vi stort sett iakttar.

Atmosfærens art og tetthet er utvilsomt av største betydning for utformingen av klodens overflateskikt. De store ytre planetenes overflater er skjult for oss av tette skylag. Det samme gjelder Venus, som ellers er meget lik Jorden i alle konstaterbare egenskaper, unntatt at solstrålingen er dobbelt så sterk og dagene i hvert fall meget lengere enn våre.

Bortsett fra Jorden selv er det bare Månen og i noen grad Mars vi kan studere overflaten på. Sammensetningen av Mars's tynne atmosfære kjenner vi ikke; den inneholder sannsynligvis vanndamp, men så lite at skyer og nedbør bare opptrer ytterst sparsomt. Solstrålingen hemmes praktisk talt ikke av atmosfæren, men dens intensitet er mindre enn halvt så stor som på Jorden, og temperaturen tilsvarende lavere. Marsoverflaten ser ut til å bestå mest av ørkenaktige strøk, men ellers er det som bekjent delte meninger om overflatedetaljene og deres tydning.

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[1] W. Kuhn und A. Rittmann: Über den Zustand des Erdinnern und seine Entstehung aus einem homogenen Urzustand. Geolog. Rundschau *32*, 215, 1941.

Forholdene på vår egen måne kjenner vi meget bedre. Solstrålingen, som har samme styrke som på Jorden, når helt usvekket overflaten og beskinner hvert punkt uavbrutt i et par uker om gangen, for så å bli borte like lenge. Dette fører til voldsomme temperatursvingninger. Månens overflateformer er kjent i fineste detalj, riktignok bare på den ene halvkulen; store deler av Jordens overflate er ikke på langt nær så godt kartlagt som Månen: detaljer med noen få hundre meters tverrmål er synlige med våre beste kikkerter. Noen forandring i disse detaljene er ikke konstatert så lenge Månen har vært overvåket med gode kikkerter. Vulkanske utbrudd og dannelsen av større forkastninger, fjellskred o. l. ville sikkert være blitt sett, og vi må derfor anse Månen som en „død" klode. Manglen på atmosfære gjør også at denudasjon og transport av løsmateriale ikke kan foregå, og de overflateformene vi nå ser på Månen må derfor stort sett ha vært slik siden de ble til. Det er blitt bygget opp høye og bratte fjellformasjoner i flere generasjoner. Over store strøk har vi eldre og yngre fjellformasjoner innfiltret i hverandre uten at de eldre er blitt nevneverdig forstyrret. Andre steder („marene") er eldre fjellformasjoner blitt delvis begravet i materiale med noenlunde jevn overflate, kanskje lavastrømmer. Fjellformasjonene på Månen er som bekjent meget karakteristiske. Sirkelformen går igjen overalt, fra de små „kraterne" til de store ringfjellene og ringvollene, og selv de såkalte fjellkjedene („Alpene", „Kaukasus", o. a.) innrammer omtrent i sirkelform de store „marene". Vi kan vanskelig forklare oss disse ringene uten å tenke oss en utslynging av stoff fra ringenes sentrer, vel som følge av vulkanske eksplosjoner, kanskje også nedslag av meteoritter. Mange av ringene har riktignok en radius på mange norske mil, „marene" enda meget mer, men i Månens svake gravitasjonsfelt (Tabell 1, m/r^2) og uten luftmotstand skal det ikke abnormt kraftige eksplosjoner til for å slynge stoff både så langt og lenger bort. Hvis det er riktig at månefjellene er blitt til på en slik måte kan de ikke bestå av fast fjell, men må være oppbygget av skarpkantet (uforvitret) løsmateriale. Observasjoner av polarisasjonsgraden av det reflekterte sollys fra Månens overflate og av temperaturfallet under måneformørkelser tyder også på at Månen overveiende er dekket av materiale med egenskaper som vulkansk aske og pimpstein. Til tross for at ringfjellene er noe ujevnt fordelt får vi inntrykk av at den faste fjellgrunnen stort sett er av samme karakter over hele Månens overflate. I hvert fall finner vi ikke noe tilsvarende til det mest påfallende trekk ved Jordens overflate, motsetningen mellom kontinenter og oseanbekkener. Her spiller utvilsomt Jordens atmosfære og hydrosfære — og dermed i siste instans dens masse og radius — en viktig rolle.

I kontinentene er på et forholdsvis begrenset areal opphopet størstedelen av jordskorpens spesifikt lette bergarter, „sial"-skorpens bergarter; det er derfor kontinentene raker flere kilometer opp over oseanbunnenes middelnivå. Oppbyggingen av kontinentene skyldes vel i første rekke gjentatte orogeneser; vi finner jo i deres fjellgrunn — hvor den er tilstrekkelig blottlagt — metamorfe bergartkomplekser som må tydes som rester av gamle fjellkjeder, og vi ser også at yngre fjellkjeder kan oppstå langs randen av eldre kontinenter, som på den måten får en tilvekst. Opprinnelsen til fjellkjedenes og dermed kontinentenes spesifikt lette bergarter er for en vesentlig del, kanskje overveiende, geosynklinalsedimenter, flysch- og molassesedimenter, og disse kan ikke bli til i et vakuum, som på Månen, men bare ved atmosfærilienes nedbrytende, oppløsende, transporterende, sorterende virksomhet. Orogenesen er naturligvis i seg selv en endogen prosess, som også ledsages av magmatisk aktivitet — bl. a. differensiasion — men i yngre fjellkjeder er det mest sedimenter som er stablet opp til store mektigheter, og i gamle neddenuderte fjellkjeder finner vi store masser av sure gneisbergarter og graniter som mer eller mindre overbevisende tolkes som metamorfe og metasomatisk forandrete sedimenter. Meget tyder på at en „sial“-skorpe som Jordens ikke ville komme til utvikling på en planet som mangler atmosfære, — eller som har en atmosfære av vesentlig annen karakter enn Jordens — selv om en må regne med noen virkning av magmatiske differensiasjonsprosesser.

Etter foredraget framkom bemerkninger av Størmer, Broch, Sæther, Münster Strøm, Chr. Oftedahl og foredragsholderen.

EKSKURSJONER

Ekskursjon 29, 31. mai 1945, til „Bærumskalderaen“ under ledelse av Egil Sæther. 10 deltagere.

Tilføyelser om tidligere ekskursjoner.
Ekskursjon 27, (omtalt b. 23, p. 239) var til Vettakoll-området. Foruten Holtedahl's framvising av eruptivgeologien demonstrerte G. Holmsen de kvartærgeologiske forhold ved den marine grense i Skådalen.

Ekskursjon 28, (omtalt b. 24, p. 122) 7 juni 1944 til Lunner på Hadeland bør minnes for den „hold-up“ ekskursjonens deltagere var gjenstand for fra det såkalte statspolitis side. Det var tre mann, rene bandittyper, som deltok. To dekket deltagerne med sine maskinpistoler, som de truet med å anvende. Så forvåket og overnervøs som den ene stod med fingeren på avtrekkeren var situasjonen ikke uten risiko, ytterligere opphisset som han ble under ordskiftet med ekskursjonens deltagere. Sluttelig lyktes det å klargjøre at deltagerne virkelig var Geologisk Forenings medlemmer på ekskursjon, hvoretter de fikk fortsette.

Lov for Norsk geologisk forening.

§ 1. Norsk geologisk forenings opgave er å bidra til utvikling av geologisk kunnskap i teoretisk og praktisk retning. Foreningen vil arbeide for dette ved møter med diskusjon og om mulig også på andre måter. Foreningen utgir et geologisk tidsskrift.

§ 2. Foreningen har et styre bestående av formann, sekretær, redaktør og to andre medlemmer samt to varamenn.

Styret kan, om det måtte finne det påkrevet, ansette en lønnet regnskapsfører for et år om gangen. Lønnen må godkjennes av generalforsamlingen.

§ 3. Formannen sammenkaller og leder foreningens møter. I tilfelle av forfall fungerer som stedfortreder et av styrets øvrige medlemmer.

§ 4. Formannen og sekretæren forvalter foreningens midler.

§ 5. Foreningens tidsskrift, Norsk geologisk tidsskrift, redigeres av styret med den valgte redaktør som hovedredaktør. Det utsendes såvidt mulig regelmessig med 4 hefter om året.

§ 6. Foreningen holder generalforsamling hvert år innen utgangen av februar måned. Det holdes hvert år 6 ordinære møter på en av styret fastsatt dag i månedene februar, mars, april, mai, november og desember.

§ 7. Sekretæren har i generalforsamlingen å framlegge melding for det foregående år og regnskapet i revidert stand.

På generalforsamlingen velges styre med 2 varamenn samt 2 revisorer med varamenn for det kommende år. Den samme formann kan ikke velges mer enn 2 år på rad.

Valget av styre foregår således: Først velges formann, sekretær og redaktør særskilt. Deretter velges 4 andre medlemmer: de to, som har flest stemmer, blir styremedlemmer, de to andre varamenn.

Skriftlig avstemning brukes ved valg av styret, og hvis noget medlem framsetter krav derom, også ved andre avgjørelser. I tilfelle av stemmelikhet avgjøres valget ved loddtrekking. Hvert personlig medlem har én stemme.

§ 8. Medlemskontingenten er 10 kr. pr. år. Man kan bli medlem for livstid ved å innbetale 100 kr. en gang for alle eller 30 kr. pr. år i 4 år i trekk. Dog kan medlemmer, som har betalt kontingent i minst 20 år, søke styret om å bli livsvarig medlem ved å innbetale 50 kr., og medlemmer, som har betalt i 15 år, ved innbetaling av 75 kr. Kontingenten for de livsvarige medlemmer opbevares som et fond, hvis midler styret i samråd med revisorene anbringer i gode verdier. Styret treffer bestemmelse om bruken av fondets avkastning.

§ 9. Forslag om å opta nye medlemmer innsendes til styret og må være undertegnet av 2 medlemmer. Styret forelegger på førstkommende ordinære møte forslaget for foreningen, som treffer avgjørelse om innvalget. For innvalg kreves minst $^2/_3$ av de avgivne stemmer. Etter forslag av styret kan geologer utenfor Skandinavia innvelges som foreningens korresponderende medlemmer. Det er adgang for alle, også for bedrifter, å tegne seg som støttende medlemmer av foreningen ved årlige bidrag eller bidrag en gang for alle.

§ 10. Forslag til forandring i foreningens lover behandles og avgjøres i generalforsamlingen, når forslaget er kunngjort for medlemmene minst 10 dager forut. Ved avstemningen utkreves $^2/_3$ av de avgivne stemmer for at beslutningen kan være lov. Forslagene vedtas eller forkastes punktvis.

Statuter for Berg- og steinindustriens gavefond til støtte for Norsk geologisk tidsskrift.

§ 1. Berg- og steinindustriens gavefond til støtte for Norsk geologisk tidsskrift er stiftet 1943 på grunnlag av gaver fra en rekke personer og bedrifter, vesentlig innen berg- og steinindustrien. Fondets grunnkapital er kr. 20 000,00.

§ 2. Fondet forvaltes av en komité valgt av Norsk geologisk forening. Komitéen anbringer fondets midler. Fondets kapital må ikke røres. Det kan økes ved avsetning av den årlige avkastning, ved gaver eller på annen måte. Komitéen skal avlegge årsmelding og regnskap, revidert av foreningens revisorer, for generalforsamlingen.

§ 3. Komitéen skal bestå av 3 medlemmer som velges av generalforsamlingen for 3 år.

§ 4. Forandringer i disse statuter kan bare gjøres på samme måte som fastsatt i lov for Norsk geologisk forening § 10 for forandringer i samme lov.

NORSK GEOLOGISK FORENING

MEDLEMSLISTE

à jour 18. oktober 1945.

*: livsvarig medlem.

Tallet i parentes er innvalgsåret.
(S): stifter (18. februar 1905).

*Adamson, Olge, J. dr. Mineralogisk institutt, Blindern, V. Aker. (1945).
Ahlmann, Hans W: son, professor. Stockholms högskola. (1919).
Alsgaard, P., ingeniør. Herøya, pr. Porsgrunn. (1945).
Asklund, Bror, statsgeolog, Sveriges geologiska undersökning, Stockholm 50. (1938).
*Bache, Laura, lektor. Nordengvegen 14, Røa, V. Aker. (1931).
*Backlund, Helge, professor. Universitetet, Uppsala. (1918).
*Balk, Robert, professor. Depertement of geology, Mount Holyoke College, South Hadley, Massachusetts, U. S. A. (1933).
*Barth, Tom., professor. Mineralogisk institut, Universitetet, Blindern, V. Aker. (1921).
Bergersen, Birger, professor. Tannlægehøgskolen, Oslo. (1921).
Bertheau-Hansen, Chr., bergingeniør. Rikstrygdeverket, Solligata 13, Oslo. (1942).
*Bjørlykke, Harald, dosent. Norges tekniske høgskole, Trondheim. (1923).
*Broch, Olaf Anton, statsgeolog. Norges geologiske undersøkelse, Oslo. (1920).
Bruun, Br., ingeniør. Geologisk museum, Oslo. (1938).
*Braastad, Johan, dr. Postboks 808, Oslo. (1913).
Bugge, Arne, statsgeolog. Norges geologiske undersøkelse, Oslo. (1914).
*Bugge, Carl, direktør, Norges geologiske undersøkelse, Oslo. (S).
*Bugge, Jens, universitetsstipendiat, Hegdehaugsvegen 5 b, Oslo. (1940).
*Callisen, Karen, museumsinspektør. Mineralogisk og geologisk Museum, Østervoldgade 7, København K. (1917).
Carlson, Fredrik, overingeniör. Sandelsgatan 35^{V}, Stockholm. (1919).
*Carstens, C. W., bergingeniør. Saxenborg, Trondheim. (1911).
Dahl, Erling, ingeniør. Heistad st., Eidanger. (1945).
Dahll, Jeanette, stud. real. Valberg, Kragerø. (1943).
Dal, Adolf, lektor. Middelthuns gate 14, Oslo. (1905).
*Damm, C. O. B., bergmester. Nes på Hedmark. (1905).
*Danielsen, D. A., rektor. Hornnes, Setesdal. (1905).
Dietrichson, Brynjulf, bergingeniør. Geologisk museum, Oslo. (1917).
Donnay, J. D. H., professor. Faculté des sciences de l'Université Laval, Boulevard de l'entente, Québec, Kanada. (1937).
Dons, Johannes A., stud. real.. Slyngvegen 14, Smestad, V. Aker. (1941).

*Eckermann, Harry von, professor. Edeby pr. Sparreholm, Sverige. (1937).
Egeberg, F. P., bergingeniør, Rådhusgata 5 b, Oslo. (1939).
*Eskola, Pentti, professor. Min. inst., Universitetet, Helsingfors. (1919).
*Falkenberg, Otto, ingeniør. Priorvegen 7, Smestad, V. Aker. (1914).
*Foslie, Steinar, statsgeolog. Norges geologiske undersøkelse, Oslo. (1911).
*Fægri, Knut. dr. Bergens museum, Bergen. (1935).
*Føyn, Sven, lektor. Kongsberg høgere almenskole, Kongsberg. (1932).
Garthus, Torbjørn, lærar. Garthus, Valdres. (1943).
Gavelin, Axel, overdirektör. Sveriges geologiska undersökning, Stockholm 50. (1920).
*Gjelsvik, Tore, cand. real. Mineralogisk institutt, Blindern, V. Aker. (1945).
Gjessing, Leiv, cand. mag. Dikemark, Asker. (1945).
*Gleditsch, Chr., stud. real. Boks 18, Skøyen ved Oslo. (1941).
Goldschmidt, V. M., professor. Geologisk museum, Oslo. (1906).
Graff, O. F., bergingeniør. Visnes, Karmøy. (1939).
Grenness, Johannes, ingeniør. Geologisk museum, Oslo 45. (1945).
*Grip, Erland, fil. lic.. Boliden, Sverige. (1938).
*Grønlie, Arne, lektor. Orkdal landsgymnas, Orkanger. (1943).
*Grønlie, O. T., rektor. Heimdal st., Sør-Trøndelag. (1909).
*Hawkes, Leonard, geologist. Bedford college, Regent's Park, London N. W. 1. (1915).
*Heber, Gustav, advokat. Oscars gate 49, Oslo. (1938).
*Heber, Gustav, bygningsingeniør. Norsk teknisk byggekontroll, Oscars gate 46 b, Oslo. (1942).
*Heintz, Anatol, professor. Geologisk museum, Oslo. (1926).
Heltzen, Anders M., stud ing.. Borgundveien 41, Ålesund. (1941).
*Helverschou, Julius, bergingeniør, disponent. Norsk diamantborings A/S, Dronningens gate 2, Oslo. (1918).
*Helverschou, Kaare, bergstuderende. Villavegen 9, Frøen, V. Aker. (1943).
*Henningsmoen, G., cand. mag. c/o Lundblad, Observatorie Terasse 5 b, Oslo. (1941).
Hesse, P. A. Jürgen, Ing.-geolog. (1936).
*Holmsen, Andreas, bergmester. Bestun, Ullern. (S).
Holmsen, Bjarne, direktør. Kjøpsvik. (1941).
*Holmsen, Gunnar, statsgeolog. Norges geologiske undersøkelse, Oslo. (1908).
*Holmsen, Per, assistentgeolog. Norges geologiske undersøkelse, Oslo, (1940).
*Holtedahl, Olaf, professor. Geologisk institutt, Blindern, V. Aker. (1908).
*Horn, Gunnar, bergingeniør. Svalbardkontoret, Observatoriegata 1, Oslo (1917).
*Hødal, Johanne, cand. real. Bergens museum. Bergen. (1943).
*Høegh, Ove Arbo, konservator. Trondheims museum, Trondheim. (1924).
Høeg-Omdal, S. K., konsulent. Utgarveien 12, Jar pr. Oslo. (1945).
*Isachsen, Fridtjov, dosent. Geografisk institutt, Blindern, V. Aker. (1929).

Jacobsen, Ruth Clementine, cand. real. Geologisk institutt, N. T. H., Tr.heim. (1941.)
Jensen, Einar, assistent. Kjemisk institutt, Blindern, V. Aker. (1939).
Johnson Høst, Mimi, læge. Lillestrøm. (1913).
*Kiil, Erling, bergingeniør. Brantenborgvegen 2, Slemdal, V. Aker. (1922).
Klüver, Emil, ingeniør. Statens toll-laboratorium, Oslo. (1926).
*Kolderup, Niels-Henr., professor. Bergens museum, Bergen. (1919).
Kollerud, Martha, cand. mag.. Trondheimsvegen 10, Oslo. (1919).
*Koren, Wilhelm, direktør, Buhemba mines, p. o. Musoma, Tanganyika Territory, East Africa. (1916).
Kristoffersen, Kr., konservator. Geologisk museum, Oslo. (1941).
Kvale, Anders, amanuensis. Bergens museum, Bergen. (1936).
Kvalheim, Aslak, ingeniør. Geologisk museum, Oslo 45. (1944).
Landergren, Sture, laborator. Riksmuseet, Stockholm 50. (1943).
Landmark, Einar, direktør. Rådhusgata 5 b, Oslo. (1941).
Landmark, Kaare, mag. scient.. Bergens museum, Bergen. (1943).
Lange, Fr., lektor. Ringstabekkveien 102, Bekkestua, Bærum. (1938).
*Larsson, Tollef, cand. mag.. Vidarsgate 13, Oslo. (1945).
*Lenander, N. E., direktør. Nordlandsgatan 7—9, Stockholm. (1914).
*Lindley, Henry W., dr. Berlin-Lichterfelde, Weddingen Weg 56. (1928).
Lund, Kjell, direktør. A/S Sulitjelma gruver, Fr. Nansens pl. 6, Oslo. (1938).
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Lund, Worm Hirsch, direktør. Jacob Fayes veg 2, Bygdøy ved Oslo. (1941).
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77 livsvarige medlemmer.
78 årsbetalende medlemmer.

155 medlemmer.

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